

Distribution of Crustal Types in Canada Basin, Arctic Ocean

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Abstract

Seismic velocities determined from 70 sonobuoys widely distributed in Canada Basin were used to discriminate crustal types. Velocities of oceanic layer 3 (6.7 -7.1 km/s), transitional (7.2-7.6 km/s) and continental crust (5.5-6.6 km/s) were used to distinguish crustal types. Potential field data supports the distribution of oceanic crust as a polygon with maximum dimensions of ~340 km (east-west) by ~590 km (north-south) and identification of the ocean-continent boundary (OCB). Paired magnetic anomalies are associated only with crust that has oceanic velocities. Furthermore, the interpreted top of oceanic crust on seismic reflection profiles is more irregular and sometimes shallower than adjacent transitional crust.

The northern segment of the narrow Canada Basin Gravity Low (CBGL), often interpreted as a spreading centre, bisects this zone of oceanic crust and coincides with the location of a prominent valley in seismic reflection profiles. Data coverage near the southern segment of CBGL is sparse. Velocities typical of transitional crust are determined east of it. Extension in this region, close to the inferred pole of rotation, may have been a magmatic. Offshore Alaska is a wide zone of thinned continental crust up to 300 km across. Published longer offset refraction experiments in the Basin confirm the depth to Moho and the lack of oceanic layer 3 velocities. Further north, towards Alpha Ridge and along Northwind Ridge, transitional crust is interpreted to be underplated or intruded by magmatism related to the emplacement of the High Arctic Large Igneous Province (HALIP). Although a rotational plate tectonic model is consistent with the extent of the conjugate magnetic anomalies that occupy only a portion of Canada Basin, it does not explain the asymmetrical configuration of the oceanic crust in the deep water portion of Canada Basin, and the unequal distribution of transitional and continental crust around the basin.

1 Introduction

While most of the world's ocean basins have reliable dates to constrain their tectonic history, the remote and ice-covered Canada Basin in the Arctic Ocean is enigmatic. The lack of clear magnetic anomalies and the paucity of seismic profiles have allowed controversy to flourish on how Canada Basin formed. Hypotheses about the nature of the crust underlying Canada Basin have ranged from oceanization of the continental crust (Shatsky, 1935) to Paleozoic or Cretaceous ocean crust from the Pacific (Churkin and Trexler, 1980) to Mesozoic oceanic crust created through diverse (and often conflicting) opening directions (Lawver and Scotese, 1990; Grantz et al., 2011; Døssing et al., 2013). On-land geological similarities (Embry, 1990) and recent detrital zircon studies (Gottlieb et al., 2014) permit

the hypothesis that the Canadian Arctic Islands and northern Alaska were once contiguous but became separated by the seafloor spreading that created Canada Basin. However, the models remain uncertain because of overlaps and inconsistencies in the reconstruction models (e.g., Lawver and Scotese, 1990; Lawver et al., 2002; Lane, 1997; Lane et al., 2015). One way to help resolve conflicts among these many models and reconstructions is to identify crustal types and their distributions within Canada Basin.

Velocity characteristics of rocks can be used to distinguish and map distinct lithologic and crustal types, such as oceanic crust (White et al., 1992). Between 2007 and 2011, Canadian and American collaborators acquired wide angle reflection and refraction seismic data from 152 sonobuoys in the Arctic Ocean primarily to study sedimentary velocities and depths and, where data quality enabled longer source-to-receiver offsets to be analyzed, to also study the deeper basement and crustal components. Of these, 87 sonobuoys are within Canada Basin and 70 are of sufficient quality to be used in this study. Offsets up to 40 km allowed the velocities of crustal arrivals to be measured. In contrast with older studies, the sonobuoy data were acquired with modern GPS navigation and digital recording technology. These seismic data, because of their regional distribution, provide the opportunity for 3D synthesis of velocities and crustal types on a basin-wide scale. This is a major difference from data typically collected on high-resolution 2D transects of continental margins (e.g. Chian et al., 1999; White et al., 2008; Funck et al., 2011). This paper presents the velocity synthesis of crustal layers, particularly the distribution and nature of crustal types based on wide angle reflection/refraction sections, potential field data, and coincident vertical incidence seismic reflection profiles. The newly defined distribution of crustal types adds a new constraint on the formation of Canada Basin.

2 Geological Framework

Canada Basin is a small ocean basin bordered on the south and east by the Alaskan and Beaufort Sea and Canadian Arctic Islands continental margins respectively, on the west by Northwind Ridge of the Chukchi Borderland, and on the north by Alpha Ridge (Fig. 1). The Alaskan, Beaufort and Canadian Arctic Islands margins are sediment covered passive continental margins, although tectonic deformation associated with the northeast extent of the Brooks Range deforms the southeastern Alaskan margin (Helwig et al., 2011; Houseknecht and Bird, 2011). At the southeast corner of the basin, the Mackenzie Delta/fan system extends for ~ 350 km onto the seafloor to ~3800 m water depth. The large structures bounding Canada Basin are diverse. Northwind Ridge is a steep, linear escarpment rising ~2800 m from the abyssal sea floor. The southern edge of Alpha Ridge has irregular and occasionally steep morphology. Between Northwind Ridge and Alpha Ridge lies the Nautilus Basin. Stefansson Basin forms a similar semi-enclosed basin where the Canadian margin joins Alpha Ridge. This paper is restricted to an interpretation of the crust in the region between the northern tip of Northwind Ridge and the Canadian polar margin, south to Alaskan margin (Fig. 1).

The study of passive continental margins is relevant to understanding Canada Basin, because the transition from continental to oceanic crust involves distinct changes in velocity structure, crustal thickness, and reflection character, depending upon whether the passive margin is magma-rich (i.e., volcanic) or magma-poor (i.e., non- volcanic). A schematic cross-section (Fig. 2) shows the main characteristics and distinguishing features of these end member crustal types. Commonly, a transition zone is interpreted to separate oceanic from adjacent crustal types. The nomenclature for this zone has been diverse, ranging from the simple (transition zone, e.g., Whitmarsh et al., 1993; Pickup et al., 1996) to the complex (e.g., multiple domains of Sutra and Manatschal, 2012; Peron-Pinvidic et al., 2013). Along some continental margins, this zone of transition is also sometimes called hyperextended

(Whitmarsh et al., 2001; Sutra and Manatschal, 2012). The work by Lundin and Dore (2011) notes confusion in using the term hyperextension because the term also refers to extended Early Cretaceous rift basins that predate the early Cenozoic breakup of the North Atlantic, i.e., the origin of the hyperextended crust is not necessarily from the final breakup process. Because the analysis in this paper focuses on the extent of oceanic crust, rather than the full transects across the continental margins, and in order to avoid confusion, we use the simplified terminology for thinned continental crust, transitional crust, and oceanic crust (e.g. Chian et al., 1995; Pickup et al., 1996).

Oceanic crust is well known and remarkably consistent in its thickness, lithology, and geochemistry in most of the world's oceans (e.g., White et al., 1992). White et al. (1992) pointed out that one of the triumphs of early marine geophysics was recognizing the oceanic crustal structure being basically the same everywhere. The two crustal layers beneath the sedimentary section have distinct and characteristic velocities: layer 2 has variable velocities (4.5 – 6.6 km/s), a large near-surface velocity gradient (~ 1.0 /s), and a thickness of ~ 2.1 km; layer 3 has more consistent velocities (6.7 km/s but no higher than 7.1 km/s), a smaller velocity gradient, and a thickness about twice that of layer 2. Total average thickness of normal oceanic crust for the North Atlantic is 7.08 ± 0.78 km (White et al., 1992). These values are consistent where oceanic spreading rates are greater than 20 mm/yr. Oceanic crust is generally thinner along fracture zones and at slower rates of spreading, where melt is presumed to be limited (Reid and Jackson, 1981; White et al., 1992).

The margins around the southern and central Canada Basin are inferred to be magma-poor (nonvolcanic) passive margins based on the interpreted lack of rift-related volcanic activity in this region (Grantz et al., 1979; Embry and Dixon, 1990; Hall, 1990; Sherwood, 1992; Houseknecht and Bird, 2011). Off the southern Labrador margin, a similar magma-poor margin, the velocity of the continental crust is

5.0-6.6 km/s (Chian et al., 1999). By analogy, the margins around southern Canada Basin are characterized by velocities associated with continental crust or thinned continental crust, i.e., velocities that are no higher than ~6.6 km/s and thicknesses that could range from ~35 to <10 km (Stephenson et al., 1994; Helwig et al., 2011).

In contrast, the northern Canadian polar margin near Alpha Ridge is interpreted as a magma-rich (volcanic) rifted margin having thinned and intruded continental crust underlain by a high velocity layer of 7.5-7.6 km/s interpreted to be volcanic (Funck et al., 2011). Tholeiitic basalt flows and sills broadly distributed across the Arctic are known on Ellesmere, Axel Heiberg, north Greenland and other polar islands (Estrada and Henjes-Kunst, 2004; Tegner et al., 2011; Evenchick et al., 2015). Therefore, the northern Canada Basin might be expected to contain thinned continental crust that changes into a zone with features similar to magma-rich passive margins, i.e., seaward-dipping reflections associated with extrusive subaerial and submarine basalt flows and a high-velocity intruded and underplated lower crustal layer in which velocities are 7.4 km/s or higher but less than upper mantle velocities of ~8.0 km/s (e.g. White et al., 2008).

High velocities are also associated with some magma-poor continental margins where serpentinized upper mantle directly underlies the post-rift sedimentary section (Whitmarsh et al., 1993; 2001; Chian et al., 1995; Pickup et al., 1996), similar to results off Iberia (e.g., Whitmarsh et al., 2000). This zone has refraction velocities of 4.5-7.0 km/s from a 2-4 km thick upper layer overlying a lower layer with a velocity of up to 7.6 km/s. These velocity models provide the analogs for which the velocities measured in Canada Basin can be associated with different crustal types.

3 Methods

3.1 Data acquisition and processing

The sound source for the wide angle reflection/refraction data was an airgun array consisting of 2x500 and 1x150 in³ SercelTM G-guns totaling 1150 in³. The airgun array was fired at a 14-20 s interval for an average trace spacing of 20-30 m. The receivers were UltrasonicTM expendable sonobuoys with a single hydrophone deployed at 60 m depth (or occasionally at 300 m depth). The seismic impulses were radio telemetered back to the stacked YageiTM array on the ship. The sonobuoy signals were digitized onboard ship and stored in SEG Y format (e.g. Mosher et al., 2013). All sonobuoy profiles and their interpretations are documented in an atlas (Chian and Lebedeva-Ivanova, 2015).

Expendable sonobuoys do not have internal navigation systems; only the deployment position is known. The direct arrival time used a water velocity profile for the sonobuoy hydrophone at 60 m depth (Lebedeva-Ivanova and Lizaralde, 2011) to calculate the source to receiver offsets. Data processing of sonobuoy wide angle reflection/refraction profiles includes a static time correction for airgun delay, digitizing arrival time T (sec) of direct water wave for all traces, and assigning offsets (Chian and Lebedeva-Ivanova, 2015). The refractions recorded in the sonobuoys traveled nearly horizontally in corresponding layers, producing direct evidence for velocity measurements in sedimentary layers and below. Nearly all sonobuoys in the central and southern Canada Basin had up to a dozen refractions observed at source-receiver offsets of 9-40 km. Since recordings spanned a wide range of frequencies, final modeling used seismic data filtered at 2-50Hz (in some cases non-filtered).

3.2 Modeling procedures

Forward modeling was accomplished using the RAYINVR program (Zelt and Smith 1992, Zelt and Forsyth 1994) and the in-house software of SeisWide (Chian and Lebedeva-Ivanova, 2015). Both the sonobuoy

data and its coincident reflection profile (Chian and Lebedeva-Ivanova, 2015) were first geometrically aligned in distance trace by trace. Major horizons on the reflection profile were digitized to generate a preliminary model in the time domain. Each layer's velocity (V) and gradient should in general match the corresponding refraction slope of the sonobuoy wide-angle data by $V = dX/dT$, which can be easily measured. This time-domain model was converted to depth for raytracing using RAYINVR, and the resulting travel-time curves were compared to the observed phases on the sonobuoy data. Such comparison was greatly enhanced with NMO-reduced sonobuoy data (Chian and Lebedeva-Ivanova, 2015), especially when coincident reflection profile were available. The NMO-reduction aimed to transform the data such that each wide-angle reflected phase at zero-offset matched exactly with the reflection data. We found that, at the resolution of our experiment, whenever refracted phases were properly modeled, we saw nearly perfect modeling for the corresponding wide-angle reflections. For layers on reflection profile that were not associated with any sonobuoy refractions, a greater vertical velocity gradient usually existed, which were modelled by matching the curvatures of the corresponding sonobuoy wide-angle reflections. In general, a few perturbations of a layer velocity were usually sufficient to arrive at a satisfactory match of the corresponding refraction phase. The process was used for all layers. Since this method worked in the two-way time domain, only perturbations in velocity (not time/depth) were needed because digitized reflection horizons were fixed during the model perturbations. Note that in the sonobuoy profiles, the slopes of crustal refractions at farther offsets can be easily identified where they became first arrivals, not entangled with reverberated and clustered reflective energy. Seismic anisotropy was not considered due to its complexity, lack of sufficient ray coverage in the relatively large study area.

3.3 Error Analysis

Error analysis of the sonobuoy data for 5 main seismic refraction phases and upper mantle reflection phase (PmP) is summarized in Table 1. When picking each phase, every trace that contains visible amplitudes was picked. The calculated uncertainties of the final model depend on the quality of the fit between the calculated and the observed model. Two indicators of the reliability of the model are the number of data points used to constrain the parameters of the layers, and the (root-mean-square) misfit (Trms) between the calculated and observed travel times. Another indicator is the pick error bar of 40-200 ms assigned by visual inspection of the data, which was used to compute the normalized X^2 misfit. In general a good fit was achieved partly due to the flat seafloor and low relief on the sedimentary /crustal layers, relatively short offsets (<35-40 km), and the low ray coverage (no rays intersecting each other). Because the sonobuoys do not overlap, the X^2 misfit compiled in Table 1 was better than some other studies using overlapping seismometers such as Funck et al. (2011).

Table 1: Error analysis of selected sonobuoy velocities in Canada Basin (south of Alpha Ridge Igneous Province), based on selected travel-time picks in continental (type 2), oceanic (type 3), and transitional (type 4) zones. Type 1 includes other (unassigned) observed upper crustal picks that are used in this study. Trms gives the root-mean-square misfit between calculated and picked travel times.

Crustal Type	Interval Velocity (km/s)	No. of SBs	No. of Picks (total: 18704)	Trms (s)	Normalized X^2	Velocity Error Bound (km/s)	Depth Error Bound (km)
1. Upper (Unassigned)	$4.7 \leq V \leq 5.4$	6	1013	0.045	0.416	± 0.2	± 0.5
2. Continental	$5.5 \leq V \leq 6.6$	7	1154	0.041	0.259	± 0.2	± 1.0

3. Oceanic	$6.7 \leq V \leq 7.1$	24	4092	0.035	0.285	± 0.2	± 1.0
4. Transitional	$7.2 \leq V \leq 7.6$	28	5819	0.036	0.242	± 0.2	± 0.5
Upper mantle	$V \geq 7.7$	13	1619	0.040	0.233	± 0.1	± 1.0
Moho	(PmP)	22	5007	0.045	0.301		

To estimate the resolution of velocity models, the range of permissible variations in the models was obtained by perturbing velocity values until the computed travel time curves cannot fit the observed phases in both its slope and time. As described in Chian and Lebedeva-Ivanova (2015), the perturbation of layer velocities was done in time domain in which we kept horizons unchanged in time based on coincident reflection profile. For the layer of basement (upper crust in Table 1), the estimated depth error bounds was a combination of errors caused by cumulative velocity errors of all overlying layers, and interpretive (or phase pick) uncertainty on the reflection profiles and on the wide-angle data. Sub-basement resolution estimates (error bounds in Table 1) were all based on wide-angle modeling as, except in very few lines in the north (e.g. Fig. 9), reflection data did not penetrate deeper than basement.

3.4 Crustal Classification

Average sedimentary thicknesses throughout Canada Basin are ~4-5 km, and are much greater than thicknesses found above ocean crust in most of the world's oceans (Divins, 2003). In the thickest sedimentary deposits on Alaskan and Beaufort Sea margins, the deepest sedimentary rocks have velocities of 4.7-4.9 km/s. Elsewhere in the basin, the sedimentary velocities generally increase from 1.8-1.9 km/s to 3.5-4.2 km/s at two way travel times of 3-4 s sub-seafloor. The base of the sedimentary section was used as the top of the crust. The top was identified either by a high-impedance reflective unconformity in the coincident multichannel seismic reflection profiles, or by the onset of velocities of

~4.3-5.0 km/s depending on location in Canada Basin. The sedimentary section is described in detail elsewhere (Shimeld et al., 2016).

Immediately below the flat-lying sedimentary horizons is a layer generally less than 1 km thick that could contain sedimentary as well as crystalline rocks. This layer consists of discontinuous dipping reflections, locally bright continuous reflections containing unconformities, or widely spaced weak reflections. The base of this unit generally forms acoustic basement, typically fading out of diffuse reflections or in places with a distinct reflection. The velocity modeled in this layer is generally in the range of 4.2-4.4 km/s. The top of this inferred basement unit is treated as the top of crust and may also represent volcanic rocks interlayered with sedimentary deposits or Proterozoic or Paleozoic sedimentary strata over basement or, in some locations, biogenic deposits. Thus, although the determination of the base of sedimentary section is clear, the onset of the crust has an estimated uncertainty of up to ± 0.5 km.

In order to derive an objective identification of crustal types, five velocity categories were assigned attributes: continental, oceanic, and transitional crusts, as well as two undifferentiated categories representing upper crust and upper mantle (Table 1 and Fig. 3). For oceanic crust there are two exceptions near the spreading center discussed in section 4.1.1. Fig. 3 shows the interpretation of crustal types based on velocity. Upper crust and upper mantle velocities were not used in the classification of most of the sonobuoys, exceptions are noted in sections 4.1.3.

In summary, we characterize thinned continental, transitional and oceanic crust from velocity ranges outlined in Table 1. Each sonobuoy record is assigned a crustal type independently on this basis. In the case where a sonobuoy recorded both continental and transitional velocities, it was categorized as transitional crust, because transitional crust can include thinned continental crust (e.g. Whitmarsh et al.,

1996). For sonobuoys with both oceanic and transitional velocities of 7.3 km/s, other data were used to interpret the crustal type (described in section 4.2.2.). The type categorization is extrapolated away from the sonobuoys locations using coincident seismic reflection profiles and regional potential field data. Finally, the potential field and bathymetry maps are used to assist in interpreting the boundaries of the crustal types.

4 Results

4.1 Sonobuoy data

Of the 87 sonobuoys examined (Fig. 1), 27 were classified as oceanic, slightly more (30) as transitional, and the fewest (13) as thinned continental crust. An additional 17 sonobuoys lacked information about crustal type altogether, either because sedimentary section was too thick for crustal arrivals to be measured (a case where crust is generally too stretched, fractured or broken such as in the Beaufort Sea), noisy traces in the far offsets obscured the deeper arrivals or the recording ended prematurely with no far offsets acquired. The velocity/depth profiles for continental, oceanic and transitional crust are shown in Figs. 3a, 3b, and 3c respectively. A sample sonobuoy record from each crustal type is presented in Figs. 4, 5, and 6. The sonobuoy names used in this paper are abbreviated from the names listed in the Chian and Lebedeva-Ivanova (2015) atlas, i.e. SB LSL 2010-1 is shortened to SB 1001.

Because of the harsh ice conditions, project objectives and equipment failures, the sonobuoys were not distributed evenly throughout Canada Basin (Fig. 1); the most consistent coverage is in the interior portions of the basin. The continental margin off the Canadian Arctic Islands lacks robust crustal information (only 2 sonobuoys); as well, there is a gap in the coverage where Northwind Ridge meets Alaskan margin. Hence the sonobuoy data set is most representative of the deep-water portions of Canada Basin.

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271 4.1.1 Oceanic Crust (27 sonobuoys)

272 Roughly thirty percent of the sonobuoys in Canada Basin (i.e. 27) have crustal velocities in the range 6.7-
 273 7.1 km/s, typical of oceanic layer 3 (Fig. 3b). Of these, 12 show mantle refractions (Pn) or less
 274 commonly Moho reflections (PmP) with velocities generally of 8.0-8.1 km/s, with a range in velocities of
 275 7.7-8.2 km/s. These mantle arrivals constrain the thickness of oceanic crust to be 5-6 km, somewhat
 276 thinner than typical oceanic crust of ~ 7 km (White et al., 1992), and two have crust as thin as 4 km.
 277 These thicknesses are consistent with formation in slow spreading mid-ocean ridge settings (Bown and
 278 White, 1994). The velocity/depth functions for these sonobuoys (Fig. 3b) favorably compare with those
 279 observed in the North Atlantic slow spreading oceanic crust.

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281 SB 0726 (Fig. 4) is representative of data from the oceanic region. The seismic reflection profile (Fig. 4a)
 282 shows a change in the amplitude, frequency and continuity of the reflections at the base of the thick
 283 flat-lying sedimentary section. The refracted arrivals at the base of this section have been interpreted to
 284 have a velocity of 4.4 km/s. Beneath this layer a velocity of 4.7 km/s is observed. This layer could
 285 represent the upper part of oceanic layer 2 and might be composed of basalt flows intermixed with
 286 sedimentary deposits. Refractions with velocities of between 5.4 and 6.3 km/s are associated with the
 287 upper layers of the crust. The above velocities are measured at the top of the layers and have high
 288 velocity gradients typical of oceanic layer 2. A clear 6.9 km/s first arrival is observed over a 25-32 km
 289 offset (Fig. 4b). At offsets of 33-39 km, a refraction with a velocity of 8.2 km/s (Pn) is observed as the
 290 first arrival. Crustal thickness is about 5.5 km.

291

292 In the central Canada Basin, the majority of the sonobuoys results indicate oceanic crust. However, SB
 293 0804 and SB 0805 do not show well developed layer 3 velocities (6.7-7.1 km/s) in the middle of the

oceanic region (Figs. 1 and 7). The crustal velocities determined for these two sonobuoys are 6.6 km/s (Fig. 3b) and 7.4 km/s (Fig. 3c) respectively. The 6.6 km/s velocity has an error bar of ± 0.2 km/s and has been plotted in the oceanic category. The 7.4 km/s velocity requires assigning it to a transitional category. Both of these are located close to the CBGL anomaly (Fig. 8) that has been interpreted to mark the fossil spreading center (Laxon and McAdoo, 1994). Velocity/depth functions lacking layer 3 have been observed in the vicinity of the ultra-slow spreading Gakkel Ridge (Jokat et al., 2003; Jokat and Schmidt-Aursch, 2007; Hermann and Jokat, 2013) and also on the oblique (and ultraslow-spreading) Knipovich Ridge (Kandilarov et al., 2010; Jokat et al., 2012). On the Knipovich Ridge, total crustal thickness is anomalously thin ~ 3.5 -4.0 km and oceanic layer 3 has velocities of 5.7-6.1 km/s and a Pn of 7.-7.5 km/s. Crustal thickness on Knipovich Ridge increases to that more typical of oceanic crust at about 40 km from the ridge axis where the first clear magnetic anomalies are observed. The relatively low velocities in layer 3 are explained by thermal expansion of the young crust, fracturing and serpentinization. White et al. (1992) also points out that in fracture zones, layer 3 is often absent. They suggest the cause for this occurrence in fracture zone is that the melt has migrated laterally away instead of being built up by multiple injections from below. As the spreading ceased at Canada Basin ridge axis, a lack of melt could have also occurred, thus layer 3 would not be well developed and may explain why these two sonobuoys are within the oceanic domain but do not have layer 3 velocities.

4.1.2 Transitional Crust (30 Sonobuoys)

There are 30 sonobuoys classified as transitional crust, with diagnostic velocities of 7.3-7.6 km/s. They are geographically distributed around the oceanic crust (Fig. 1). Of these, 26 have velocities of 7.4-7.5 km/s, and no velocities were higher than 7.5 km/s (Fig. 3c). The original definition of transitional crust has velocities in the range of 4.5-7.0 km/s underlain by velocities of 7.6 km/s that are associated with

serpentinized mantle (e.g. Chian et al., 1995; Pickup et al., 1996). High-amplitude wide angle reflections (PmP) are sometimes observed, providing constraints on crustal thickness.

SB 1034 and the coincident seismic reflection profile (Fig. 5) are representative of data from the transitional crust. The sedimentary section on the coincident vertical incident seismic reflection profile shows a thick section up to 6 km of flat-lying reflectors with a velocity of ≤ 4.5 km/s (Fig. 5a). Beneath the flat-lying horizons, a higher amplitude lower frequency hummocky reflection is observed. The layer beneath this base sedimentary horizon has a 5.3 km/s velocity and a thickness of 2.5 km, based on second arrivals. A clear first arrival refractor with a velocity of 7.4 km/s is observed at the 25-35 km offsets Fig. 5b). This refractor has strong reverberations.

In the southeast part of Canada Basin in the Beaufort Sea, a group of sonobuoys are classified as transitional (Figs. 1 and 3). These sonobuoys (SB 0710, SB 0711, SB 1025, SB 1026, SB 1027, SB 1031, SB 1032, SB1033 and SB 1034, Fig. 5) have clear 4.3-4.5 km/s first arrivals that break over to 7.4-7.5 km/s. The sedimentary section is greater than 10 km thick. Sedimentary velocities as high as 4.5 km/s and a single well-defined high-velocity layer of 7.5 km/s (e.g. SB 1034) are characteristic of the region. The 7.4-7.5 km/s refractor is sometimes accompanied by a strong reverberatory band of wide angle reflection energy, suggesting that the top of the high-velocity layer may be a sharp boundary. For models to match the curve of the deepest wide angle reflections, masked layers with velocities in the range of 5.4 km/s and 6.2 km/s are required for SB 1027 and SB 1032 respectively (Fig. 3c). It suggests that continental crust of less than a few kilometers thick exists in this region. On SB 1026 (Fig. 3c), the thick sedimentary deposits may rest directly on the 7.5 km/s layer and the crust may be absent.

North of the Alaska margin along Northwind Ridge, velocities interpreted on SB 0722, SB 1013 and SB1014 indicate that thin (continental) crustal layers may also be present (Figs. 1 and 3c). On SB 0722 a 6.1 km/s upper continental crustal velocity lies immediately above the 7.5 km/s layer. Nearby and closer to Northwind Ridge, SB 1013 (Fig.1 and 3c) has well developed base of sedimentary section arrivals and a weak 5.0 km/s refractor overlying a 7.4 km/s event and a possible weak Pn arrival that would limit the depth to Moho to 17 km. SB 1014 to the north of these sonobuoys and adjacent to Northwind Ridge has crustal velocities in the range of, 6.0-6.6 km/s overlying 7.4 km/s. SB 1015 slightly to the east records a 6.3 km/s layer overlying a 7.4 km/s refractor with an associated weak PmP arrival. This velocity range of above 7.4-7.5 km/s is compatible with an earlier result of Jackson et al. (1995) in the same area, suggesting that extremely thin, heterogeneous and possibly continental crust is a common character of this transitional zone in deep water near Northwind Ridge.

4.1.3 Thinned Continental Crust (13 sonobuoys)

Thirteen sonobuoys are classified as thinned continental crust and most have velocities of 5.5-6.6 km/s with no evidence of a higher velocity 7.2-7.6 km/s deep layer (Fig. 3a). Six of these sonobuoys show strong PmP wide angle reflections indicative of an abrupt change in velocity at the crust-mantle interface.

SB 1003 with its coincident reflection profile (Fig. 6), located 260 km north of Alaskan margin, is an example of data displaying continental velocities. On the accompanying seismic reflection record, a thick sedimentary section (>6 km) with horizontal reflections is observed. Arrivals as high as 4.7 km/s (Fig. 6a) are attributed to the sedimentary section based on comparison with the accompanying seismic reflection profile. Basement relief exists with draped high amplitude reflections filling in between these basement highs. Unfortunately, the seismic reflection profile does not have the resolution to determine

whether these highs are rotated fault blocks or merely basement highs. On the sonobuoy record (Fig. 6b), first arrivals of 6.0 km/s are observed over the 20-27 km offsets. A second arrival exhibits a 6.3 km/s event (Fig. 6b and 6b inset). Modeling shows that the rays associated with the 6.0 km/s velocity travel through the layer beneath the deepest sedimentary horizon. Beyond this range another weaker first arrival is discernible with a velocity of 8.0 km/s typical of the upper mantle Pn. Based on the above constraints, the depth to the top of the upper mantle is ~14 km and the crustal thickness is ~4 km or possibly less because the coincident reflection profile shows basement relief of ~1 km.

Offshore of Alaskan margin is the region with the densest concentration of sonobuoys with velocities typical of continental crust: SB 1001, SB 1003 (Fig. 6), SB 1005, SB 1006, SB 1007, and SB 0713, (Figs. 1 and 3). For example on SB1001 (Figs. 1 and 3a), a refraction of 4.9 km/s is observed and interpreted to be continental basement. Wide angle reflections beneath this layer indicate velocities in the range of 5.1-5.4 km/s above the clear PmP. SB 0713 and SB 1003, respectively located about 300 km and 265 km north of the 2000 m contour have thinner sedimentary layers and stronger crustal events. Hence, continental crust is interpreted to extend well into the deep-water portion of Canada Basin off this part of Alaska. However, on sonobuoys nearest the margin due to the thick sedimentary section, many of the crustal arrivals are weak and the sedimentary basement layer is obscured by the multiple.

In contrast, the area offshore of the Mackenzie Delta has only two sonobuoys (SB 1028 and SB 1030) (Figs. 1 and 3a) that have been interpreted to have continental velocities. On SB 1028 refractions with velocities of 5.5 and 6.6 km/s are measured. On SB 1030 no crustal refractions are observed but wide angle reflections from the crust exist together with a clear PmP. In order to match the curve of the crustal reflections and PmP arrivals, a masked layer with a velocity of 6.2-6.4 km/s is required for the crust. This velocity is within the range of continental crustal velocities measured by Stephenson et al.

(1994) on the southeastern Beaufort Sea continental margin. Therefore, this area of thinned continental crust is constrained to be much closer to the coast than is the case further west offshore Alaska.

In the northern Canada Basin along the Canadian Arctic Islands, only two sonobuoys are classified as continental, although the data coverage is sparse. SB 0928-1 (Figs. 1 and 3a) is located in a unique, high-relief basement depression about 130 km across (Fig. 8), interpreted as a graben structure (Hutchinson et al., 2012). This sonobuoy is in deep water ~250 km seaward of the 2500 m contour, indicating thinned continental crust here, similar to offshore Alaska, may extend significantly offshore. Offshore of southern Prince Patrick Island, SB 1020 (Fig. 1) has a complex sedimentary section with low velocity zones (Chian and Lebedeva-Ivanova, 2015). An arrival of 6.6 km/s is consistent with the presence of thin continental crust (Fig. 3c).

Two sonobuoys with continental crust are also seen near the northern end of Northwind Ridge. SB 0815 is in shallow water on the ridge has a 5.0 km/s arrival and a poorly resolved 6.0 km/s event. SB 0933-1 has several crustal velocities ranging from 4.4 to 6.6 km/s and is located near its northern tip. These sonobuoys velocities are consistent with the interpretation based on geological sampling that Northwind Ridge is a continental fragment (Brumley et al., 2015).

Upper mantle refractions (Pn) are observed on SB 1003, SB 0713 and SB 0928 with continental affinities. The Pn arrival on SB 1003 and the nearby SB 0713 constrains crustal thickness to 3-4 km and 6 km respectively. The crustal thickness on SB 0928-1 located in the northern portion of Canada Basin offshore of the Canadian Arctic Islands is 7 km. Strong PmP reflections are observed on several of the other continental sonobuoys (SB 1010, SB 1003, SB 1005, SB 1006, and SB 1007) and are indicative of a

sharp boundary at the Moho (e.g. Stein and Wysession, 2003). With estimates of crustal thickness in the range of ~3-5 km and Moho depths of 11-16 km, this continental crust is therefore extremely thinned.

4.2 Integration with Potential Field Maps

The widely spaced sonobuoys and reflection profiles provide general constraints on the distribution of oceanic, transitional and continental crust. Maps of magnetic and gravity anomalies (Fig. 7 and 8) provide independent data types to further define the spatial distribution of the crustal types. The oceanic polygon is then compared with the potential field data. The circum-Arctic magnetic anomaly map (Gaina et al., 2014) shows that Canada Basin has muted magnetic anomalies compared with the large amplitude magnetic anomalies associated with Alpha Ridge just to the north of it (Fig. 7, Saltus et al., 2011). In the central Canada Basin, the region with linear conjugate magnetic anomalies (Fig. 7) has been interpreted as evidence of seafloor spreading (e.g. Vogt et al., 1982; Grantz et al., 2011).

The Free air gravity map of Canada Basin (Fig. 8) derived from Andersen et al., (2010) is dominated by large amplitude shelf-edge paired positive-negative gravity anomalies along the edges of Canada Basin that range from 50-150 mGals (Stephenson et al., 1994). An important but smaller scale feature is the narrow low amplitude negative gravity anomaly, the Canada Basin Gravity Low (CBGL), that stretches from the Mackenzie Delta area to the northeast of Canada Basin (Fig. 8). The CBGL has two parts, a southern segment that extends towards the Beaufort Sea, and a northern part, which bisects the oceanic domain polygon, consistent with its earlier interpretation as the location of the fossil ridge axis (Laxon and McAdoo, 1994). The change in direction between the northern and southern segments of the CBGL occurs about 150 km south of the edge of the oceanic domain. Although the geometry of the northern segment of the CBGL can be associated with basement topography and an axial valley in the

oceanic domain, neither our sonobuoy nor our reflection data are of sufficient resolution to determine the character of the basement on the southern segment of the CBGL.

4.2.1 Oceanic Crust

Within the polygon that describes the OCB (Table 2), the velocity data indicate oceanic crust, sub-parallel linear magnetic anomalies occur (Fig. 7), consistent with the hypothesis that oceanic magnetic anomalies are formed by seafloor spreading (Vine and Matthews, 1963). These linear anomalies have been previously interpreted as evidence of Mesozoic seafloor spreading in Canada Basin (Taylor et al., 1981, Vogt et al., 1982; Grantz et al. 2011). The character of the inner pair of aligned circular highs is similar to those observed over oceanic crust in slow or ultra-slow spreading regimes (Dick et al., 2003, Brozena et al. 2003; Cannat et al., 2006). Visual inspection suggests that the linear magnetic anomalies (Fig. 7) could extend south of the limit of oceanic crustal mapped by velocities. SB 0713 is strategically located for constraining this boundary. The 6.6 km/s velocity that is critical for identifying continental crust is a weak arrival on the sonobuoy record and, combined with our estimated error of ± 0.2 km/s, suggest the possibility that it could be interpreted as oceanic crust of a velocity of 6.8 km/s. It is also possible that the high-amplitude magnetic anomalies characterizing the onshore-offshore magnetic D3 domain of Arctic Alaska (Saltus et al., 2011) that is postulated to have thicker and probably stronger lower crust are merging with the oceanic magnetic anomalies and making it difficult to separate the two regions.

Table 2: Navigation control points for the proposed polygon of the ocean continent boundary (OCT) in Canada Basin.

Longitude($^{\circ}$ W)	Latitude($^{\circ}$ N)
-142.17053	74.02658

-140.07223	73.93067
-136.99941	73.95883
-136.44294	75.58195
-135.85894	76.02057
-135.05174	76.81987
-139.50746	78.02146
-142.9283	78.88792
-143.51109	79.12048
-143.72758	79.20687
-146.71055	79.17853
-147.95074	78.89232
-147.17291	78.56884
-148.91948	77.22398
-147.71752	76.45767
-147.61889	76.12906
-148.9723	76.04934
-147.31766	73.8471
-147.35169	73.84555
-147.25026	73.85287
-142.17053	74.02658

456

457

458 The paired conjugate magnetic anomalies fan slightly northward until they reach about 77°N. Northward

459 on the eastern (Canadian Arctic Islands) side of the basin the outer paired anomaly changes direction

and the amplitude decreases. The outer magnetic anomaly trends northwesterly and the oceanic crustal domain narrows, consistent with the sonobuoy velocity data (Fig. 7). Northward of the paired oceanic anomalies on the western side of the basin the outer paired magnetic anomaly approaches Northwind Ridge at an angle. The high amplitude Northwind Ridge Magnetic Anomaly (NoRMA) becomes the dominant magnetic feature. The velocity data are complex in this region near the northern end of Northwind Ridge. There are about 10 sonobuoys displaying the variations: for example on SB 0901-2 a 6.6-6.8 km/s velocity overlies a 7.4 km/s, on SB 0901-3 a 6.4-6.5 km/s velocity overlies a 7.3 km/s layer, and on SB 0901-4 a 6.7-6.8 km/s layer overlies a 7.3km/s event. Oceanic crust, underplating and/or intrusion of magmatic material due to the proximity of the HALIP (Maher, 2001; Døssing et al., 2013) are possible causes of the two lower crustal velocities.

4.2.2 Transitional Crust

Lundin and Doré (2011) have pointed out the difficulty of assigning velocities in the range of 7.1 -7.7 km/s to particular rock types and that geological setting is important in assessing their origin. Transitional crust in southern Canada Basin is associated with lower amplitude amorphous magnetic anomalies of variable size and orientation. Some of this pattern may be the result of the great thickness of sedimentary section beneath the Mackenzie fan, depressing both the basement and magnetic source layers, but this pattern could also result from demagnetization of the crust similar to anomalies resulting from serpentinization (Whitmarsh et al., 1993; Bronner et al., 2011). The general similarity of these anomalies with the adjacent continent, particularly the Canadian Arctic Islands (Fig. 7), suggests that at least some of this transitional crust is associated with the rifting processes that preceded formation of oceanic crust in Canada Basin. By analogy with Iberia (Whitmarsh et al., 2001), we favor the interpretation that this area of transitional crust is serpentinized mantle underlying extremely thinned continental crust.

In contrast, an elongate high-amplitude magnetic anomaly (NoRMA) dominates the transitional crust subparallel to and just east of Northwind Ridge in the western and northwestern Canada Basin. This intense anomaly is considered part of Alpha Ridge magmatic zone (Saltus et al., 2011), the 7.3-7.7 km/s velocities may be due to the emplacement of magmatic material related to Alpha Ridge (Døssing et al., 2013). The actual velocity measured on 4 sonobuoys (SB 0812, SB 0813, SB 0901-3, and SB 1016) near the northern tip of Northwind Ridge is 7.3 km/s and on SB 1017 a velocity of between 7.2 and 7.4 km/s is noted. P-wave velocities of 7.3 km/s are unique to this region and their implications are discussed in section 5.5.

4.2.3 Thinned Continental Crust

The area of thinned continental crust in southern Canada Basin, near Alaskan margin, is another region where the magnetic data are helpful in refining crustal type. Saltus et al. (2011) defined a broad positive magnetic anomaly D3 in Alaska that could extend offshore (Fig. 7) and roughly encompass the area of thinned continental crust, as defined by lower crustal velocities ≤ 6.6 km/s. The offshore portion of the D3 domain may be a structurally competent continental crustal block that resisted breaking apart during the rifting process.

4.3 Seismic Transects

Coincident seismic reflection profiles provide the opportunity to compare the reflection characteristics with crustal types inferred from sonobuoy velocity classifications and potential field data to better understand the tectonic setting of Canada Basin. Here we use depth to basement, basement topography and reflectivity to further refine our interpretations of crustal domains. Four transects (Figs. 9-12) in Canada Basin (Figs. 1, 7, 8) illustrate oceanic crust and the adjacent crustal domains.

508

509 4.3.1 North Canada Basin Transect (Fig. 9)

510 The composite profile across north Canada Basin (Fig. 9) reaches from near the northern tip of

511 Northwind Ridge to the adjacent margin offshore of northern Prince Patrick Island (Figs. 1, 7, 8). The

512 sedimentary section is generally less than 4 km thick along this transect. Oceanic crust that meets our

513 simple velocity criteria is at its narrowest width, about 150 km, and mostly consists of a deep valley and

514 the adjacent basement highs. The shape of the valley is well defined, similar to the valley on slow- and

515 ultra-slow spreading mid-ocean ridge segments (Dick et al., 2003; Ehlers and Jokat, 2009). Its position

516 coincides with the position of the CBGL, adding an important detail in support of the interpretation that

517 the CBGL represents a fossil spreading center (Laxon and McAdoo, 1994). The sedimentary deposits in

518 the valley are at least 1 km thicker than the surrounding deposits. Our interpretation of the seismic

519 section is that this is a fossil spreading center with axial valley and rift shoulders.

520

521 The top of transitional crust at the western boundary of the oceanic domain is ~1 km deeper than the

522 rift shoulder and is highly reflective. The bright reflection (Fig. 9b inset) package (Fig. 9b and c) is offset

523 by as much as 0.5 sec in many places, as if faulted, as it rises towards Northwind Ridge. Velocities within

524 this highly reflective unit (4.4-5.6 km/s) are consistent with the presence of basaltic flows, in which the

525 offsets are either due to faulting or they were emplaced on a faulted surface. Seaward dipping

526 reflections (SDRs) are also observed. On SB 1017 on the reflection line that obliquely crosses this

527 transect, a 6.2 km/s is underlain by a 7.2 km/s velocity that indicates transitional crust. However on SB

528 0901-4, just to the north of the seismic transect, a velocity 6.8-7.0 km/s typical of layer 3 is observed on

529 the OCB. Lower crust velocities of 7.2-7.4 km/s are also observed, which is the basis for interpreting this

530 crust as transitional crust. However this region could be oceanic crust with additional high velocity

531 magmatic material from the adjacent Alpha Ridge. Near where the transect crosses Northwind Ridge,

weathered basalts interpreted to originate from basalt flows and shallow water volcanoclastic deposits were dredged in 2008 and 2009 (Mukasa et al., 2009; Brumley et al., 2015). Near and on Northwind Ridge, velocities of 6.3-6.4 km/s, are interpreted to be continental crustal velocities, which is consistent with Northwind Ridge (as part of the larger Chukchi Borderland) representing a subsided crustal block.

On the eastern side of the transect, a thicker sedimentary section with associated deeper basement (location at SB 0928-1) and irregular depths to basement with abrupt offsets is imaged (Fig. 9b). There is only one seismic profile that crosses this depression but its extent can be traced along a rectangular-shaped negative gravity anomaly approximately 60 km across and 200 km long (Fig. 8). The sub-sedimentary velocity of 6.2 km/s is interpreted as continental. The edges of this deep basin on the gravity map have a northeasterly trend and could be wrench-related if the offsets are interpreted as faults on basement. If the offset bright reflections are related to flows or sill intrusions rather than faulting, the basin may alternatively be a deeply subsided pre-rift basin. The magnetic map (Fig. 7) shows a narrow linear magnetic anomaly in this deep basin, the anomaly is about 20 km across and 80 km long. At the eastern edge of the profile a prominent basement high associated with an ovoid positive magnetic anomaly (Fig. 7) may be a volcanic edifice. If the basement depth were due to extensive stretching and subsequent capping with flows or intrusives, it might compensate for the narrowing of the oceanic crust at the axial valley on this transect.

4.3.2 Central Canada Basin Transect (Fig. 10)

The composite profile across the central part of Canada Basin (Figs. 1, 7, 8) is in a region where sedimentary units are between 4 and 7 km thick. The oceanic crust is about 340 km in width here. The profile illustrates both symmetric and asymmetric features. The symmetry is illustrated by the distinctive low-amplitude, high-relief basement reflection associated with oceanic crust surrounded by high-

amplitude, low-relief reflections in the adjacent zones of transitional crust. The asymmetry is related to the shallower depth to base of the sedimentary section on the west near Northwind Ridge in contrast to the east, where a greater sedimentary thickness has presumably depressed basement.

Oceanic crust is characterized by blocky relief with offsets between 0.5 – 2.0 km. There is little reflectivity below the basement horizon. The position of the CBGL is shown and is associated with a depression in the basement relief at the position of SB 0810. The distribution of oceanic crust is asymmetric about 200 km wide in the eastern portion and 140 km wide in the western section. On the western side of the basin near Northwind Ridge, a lower crustal velocity of 7.4 km/s (i.e., transitional crustal velocities) is observed.

The base of the sedimentary layer on top of transitional crust (Fig. 10) on either side of the oceanic domain has lower relief and in the west stronger reflectivity for about 0.5 km into the crust. Transitional crust next to oceanic crust is also slightly deeper than the central oceanic basement blocks. Transitional crust at the western end of the profile consists of an upper part that is 5-8 km thick with velocities similar to either continental or upper oceanic crust (5.0-6.6 km/s) and a lower part that is 6-7 km thick with a velocity of 7.4 km/s. On the west, the zone of the transitional crust overlaps with the high-amplitude magnetic anomaly NoRMA (Fig. 7). The transitional velocities of 7.4 km/s have been called high velocity lower crust (HVLC) and associated with underplated magmatic material similar to what would be found along a magma-rich (volcanic) continental margin for the following reasons: (a) the continuity of this anomaly with the magnetic domain of the Alpha Ridge (Saltus et al., 2011), (b) the proximity of dredged basalts on Northwind Ridge (Andronikov et al., 2008; Mukasa et al., 2009), and (c) the local presence of oceanic layer 3 velocities up to 6.8 km/s in the transitional region.

On the eastern (Canadian) side of the profile, basement in the narrow transitional crust zone is obscured by the water-bottom multiple; however, where imaged in the multichannel data at its seaward (western) edge, basement relief is subdued and contains bright reflections, similar to the reflectivity patterns of basement near Northwind Ridge. SB 1020 has a thick sedimentary section and an upper crustal velocity of 5.0 km/s. The nearby refraction data of Berry and Barr (1971) (Fig. 1) support the interpretation of continental crust near the margin. The magnetic anomaly at the eastern end of the profile is subdued compared with the western end.

4.3.3 South Canada Basin Transect (Fig. 11)

The southern transect crosses Canada Basin ~300 km south of the central transect (Figs. 1, 7, 8), in a location where the sedimentary section is several kilometers thicker (Fig. 11). Good reflection and refraction control is limited to the region west of the CBGL with SB 1025 and SB 1024 constraining velocities and depths in the thick sedimentary section along the Canadian polar continental margin. Oceanic crust is 160 km wide on either side of the low amplitude CBGL. Basement deepens west of the CBGL, but whether it is a valley cannot be easily determined because of the weak reflections at the basement surface. As in the other transects, the top of oceanic crust is blocky and weakly reflective. The oceanic basement surface is poorly imaged between the CBGL and the Canadian continental margin. At 6-7 km thick (SB 0725, SB 0726), crustal thicknesses are only slightly less than those of typical oceanic crust.

Transitional crustal velocities on this transect are 7.4-7.6 km/s (SB 1013, SB 0723 and SB 0724). On the western edge of the transect (Northwind Ridge), the top of transitional basement has bright reflections (Fig. 11c). A small sub-basin located at SB 1013 could be a synrift structure. Beneath the sedimentary layers, velocities of 4.5-5.2 km/s are interpreted to be highly extended continental crust that is 4-5 km

thick above the deep transitional (7.4 km/s) body. This transitional layer is up to 5 km thick (SB 1013) and 225 km wide. The existence of a separate Moho arrival and the location of this body adjacent to Northwind Ridge suggest that this transitional crust is underplated magmatic material. Were it serpentinized upper mantle, the 7.4 km/s event would be the top of upper mantle, nor would one expect strong, clear, separate arrivals for a PmP or Pn. Near SB 0723 and SB 0724, velocities in the upper mantle are not well enough constrained to determine whether the transitional body actually thins towards oceanic crust, although the top of the transitional body does deepen from ~12 km at SB 1013 to ~14 km at SB 0724. A sonobuoy on the eastern, Canadian margin (SB 1024) yields sedimentary velocities (4.8 km/s) to a depth of 14 km below the sea surface. The nearest offline SB 1025, approximately 90 km further south, shows a 7.4 km/s event as a strong reverberation at 11-22 km offsets, but only a weak, reverberatory first arrival at offsets of 22-45 km is recorded on SB 1025. There is no indication of a deeper Pn arrival. This transect indicates major differences in crustal thickness, refraction character, and depth between the eastern and western zones of transitional crust. Our preferred interpretation is that transitional crust on the eastern, Canadian margin is serpentinized peridotite.

4.3.4 North-South Canada Basin Transect (Fig. 12)

The final transect crosses the basin from the center of oceanic crust south to Alaskan margin (Fig. 12). This transect is located west of the CBGL (Figs. 1, 7, 8). Oceanic crust in this transect is similar to its character on the central and southern transects, and is a low-amplitude, high-relief reflection. Oceanic layer 3 crustal velocities of 6.9 - 7.0 km/s are first arrivals and well constrained in SB 0726, SB 0727, SB 0728. On SB 0804, a 6.6 km/s layer is determined on a weakly defined refraction (Chian and Lebedeva-Ivanova, 2015). SB 0805 is projected into the transect to show sedimentary velocities, but is in the fossil spreading center within the CBGL anomaly and has a transitional crust velocity (7.4 km/s), similar to other velocities measured in slow and ultra-slow spreading centers (Jokat et al., 2003; Jokat et al., 2012).

628

629 A 50 km wide zone of transitional crust is adjacent to oceanic crust (SB 1004, Fig. 12), defined by a
630 strong 7.5 km/s refraction. About 10 km to the northwest of SB 1004, SB 0715 has a transitional crustal
631 velocity of 7.4 km/s, confirming the presence of this narrow strip of transitional crust. This is the
632 narrowest transition zone measured in the basin.

633

634 Along Alaskan margin, sedimentary strata are up to 10 km thick and are constrained by the presence of
635 subhorizontal, subparallel reflections down to the first multiple reflection and by velocities (Fig. 12) that
636 are as much as 4.7-4.9 km/s (SB 1007, SB 1006, and SB 1005). On the basis of observed crustal velocities
637 that are typical of continental crust (5.0- 6.4 km/s), crustal thicknesses less than 5 km, and wide-angle
638 arrivals modelled with upper mantle velocities of 8.0 - 8.1 km/s (i.e., precluding the presence of oceanic
639 or transitional velocities), this 180-km wide region is interpreted as thinned continental crust.

640

641 In summary, the reflection data add to the picture of the different crustal zones. The top of oceanic
642 crust is typically a weak reflection of high relief compared to the strongly reflective and lower relief at
643 the base of the sedimentary section in the transitional zone. A depression or valley is associated with
644 the CBGL, consistent with the interpretation that it represents a fossil spreading center (Laxon and
645 McAdoo, 1994) with a central rift valley, typical of slow-spreading to ultra-slow spreading ridges (Dick et
646 al., 2003, Jokat et al., 2003). In general, large blocks with offsets up to 2 km define the morphology of
647 oceanic crust, which is much better imaged west of the CBGL. The edges of the transitional crust, while
648 clear on the reflection seismic data where transitional and oceanic crust meet, are not well constrained
649 along the continental margins where transitional and extended continental crust meet. This is mostly
650 because of insufficient data coverage.

651

Continental crust along the Alaskan and Canadian Beaufort margins is not often imaged in the seismic reflection data because of interference from the water bottom multiple. Where velocities do not indicate oceanic or transitional crust, continental crust is highly extended and often less than 5-10 km thick. Sonobuoys at the north end of Northwind Ridge confirm its origin as a submerged continental block. Another possible continental block is buried beneath the northeast Canada Basin (Fig. 9) where a large graben structure exists.

5 Discussion

5.1 Comparison with published long offset wide angle reflection/refraction data

Several previous regional crustal refraction experiments in Canada Basin (Fig. 1) with offsets of up to 200 km (e.g. Mair and Lyons 1981) provide an important comparison with our more detailed but shorter offset sonobuoy data, especially in determining the velocities and depths of the lower crust and mantle.

5.1.1 Near Alaskan margin

Near the central Alaskan margin in southern Canada Basin (Figs. 1 and 13), Mair and Lyons (1981) used sparsely spaced explosive shots on a cross array and recorded clear refractions. The primary velocities measured were a 4.3 km/s arrival, a weaker event of 6.6 km/s with a low gradient and a better defined 7.5-7.7 km/s layer with a high gradient increasing to 8.1-8.4 km/s out to a range of 200 km. Their velocity model is projected onto our adjacent line in Fig. 13 with sonobuoys measurements and reflection profiles. Their velocities of 4.3-4.5 km/s were interpreted as representing oceanic layer 2. The 4.4-4.6 km/s velocities that we measure can be compared to our coincident reflection profiles where a thick sedimentary section is observed. Without access to reflection profiles, they assumed that these velocities represented oceanic layer 2.

676 Beneath this layer they interpreted only a few first arrivals in the 20-40 km range to constrain the 6.6-
 677 6.8 km/s velocity. However, the geometry of their cross array shows this layer thickens towards the
 678 margin and has a low gradient. On SB 1002 near the Mair and Lyons experiment, we observed a 7.5
 679 km/s refraction associated with a strong reflection that requires a continental velocity of 5.0 km/s to
 680 model its shape. On SB 1003, the basement refraction velocity of 5.9-6.0 km/s is typical of continental
 681 crust and underlies the clear sedimentary section determined from the coincident seismic reflection
 682 profile. On nearby SB 1005 (Fig. 12), beneath the sedimentary section, velocities of 4.9, 5.3 and 6.0
 683 km/s are interpreted from a noisy record section. On SB 1006 (Fig. 12), velocities of 5.0-5.3 and 6.0-6.6
 684 km/s are weak and accompanied by a strong PmP. Mair and Lyons (1981) interpreted their 6.6-6.8 km/s
 685 as oceanic. The lower velocities on the sonobuoys, the lack of linear magnetic anomalies and the surface
 686 characteristics of basement suggest thinned continental crust.

687
 688 Mair and Lyons (1981) measured a 7.6 km/s velocity consistent with our observations of 7.5 km/s
 689 velocities on SB 1002 and SB 1004. Furthermore, they were able to determine that this layer had a high
 690 gradient based on their long offset arrivals. They interpreted this layer as either part of oceanic layer 3
 691 or serpentinized peridotite. Oceanic layer 3 does not typically have a high gradient or velocities in the
 692 range of 7.5-7.6 km/s but a serpentinized layer does (Christensen, 2004). Therefore in our model (Fig.
 693 13) this layer is interpreted as serpentinized peridotite underlying thinned continental crust consistent
 694 with the generalized model of a magma-poor margin (Fig. 2).

695
 696 The sedimentary section (Fig. 13), as interpreted from the seismic reflection profile and the velocities,
 697 thickens from SB 0726 towards the margin. The crustal layer thins from the continental margin towards
 698 the centre of the basin where it is underlain by a 7.5-7.6 km/s layer. The transition to oceanic crust (Fig.
 699 13) takes place, not at the shelf break (Mair and Lyons, 1981), but 275 km seaward of the shelf break

between SB 1003 and SB 0726. As illustrated on SB 0726, well defined layer 3 velocities of 6.9-7.1 km/s and lineated magnetic anomalies are observed. The broad magnetic anomalies near SB 1006 (the D3 zone of Saltus et al., 2011) are gradually replaced by more linear anomalies that continue seaward in the region displaying clear oceanic crustal velocities. The zone where magnetic lineations are observed, but velocities suggest the presence of continental crust, occur at the ocean-continent boundary suggesting that the continental crust has been intruded with basaltic rocks.

5.1.2 Near southern Northwind Ridge

Also in southern Canada Basin, Baggeroer and Falconer (1982) carried out a crustal refraction program near and parallel to SB 1009, SB 1010 and SB 1011 (Figs. 1 and 14). They used two types of receivers, a multichannel hydrophone L-shaped array with 600 m length on each arm and an ocean bottom seismometer. The sound source was 5 explosive charges at offsets from 39 to 73 km. They recorded a very strong event with an estimated velocity of 4.9 km/s arrival and a 7.9 km/s arrival that they interpreted as the Moho. Their experiment was located only 5 km to the east of one of our reflection profiles and about midway between SB 2009 and SB 2010. SB 1009 and SB 1010 recorded high amplitude 4.5-4.6 km/s arrivals. On SB 1011, weaker but observed velocities of 5.0 and 5.8 km/s are also seen (Chian and Lebedeva-Ivanova, 2015), similar to the 4.9 km/s velocity of Baggeroer and Falconer (1982). A velocity of 7.5 km/s is observed on the sonobuoy data breaking over as a first arrival at about 25 km offset. Baggeroer and Falconer (1982) observed a velocity of 7.7 km/s from 39.4 to 73.2 km range with the 4.9 km/s event breaking over to 7.9 km/s event at 29.1 km. The depth to the high velocity layer 7.5-7.7 km/s is predicted to be 9.8-10.7 km below the seafloor by Baggeroer and Falconer (1982), consistent with the velocity model for SB 1011.

The results of both experiments are similar but difficulties in interpreting them remain. The 4.5 km/s arrival on the sonobuoy data, when compared with the coincident reflection profile, occurs within the sedimentary section. It is puzzling why a 4.5-4.9 km/s sedimentary arrival at about 4.3 km depth within a 6 km thick sedimentary section should be so strong a refraction. Baggeroer and Falconer (1982) made an important observation that this event exhibits high attenuation with offset, suggesting a sharp gradient typical of a change at the sedimentary basement interface. They also commented that their 4.9 km/s velocity was high for a sedimentary layer and could mean that this layer was compacted sedimentary material. Baggeroer and Falconer (1982) were also concerned that no layer 3 velocities had been measured. Based on spectral analysis of the array data, layer 3 was thin (less than 2 km thick) or absent. The location of their experiment coincides with the region of thinned continental crust extending north from Alaskan margin (SB 1001 to SB 1005). In our more spatially distributed and shot-intensive sonobuoy data, velocity analysis did not reveal a typical layer 3 oceanic velocity. Our favored interpretation is a thick sedimentary section overlying thin continental crust with a velocity of 5.0-5.8 km/s, with a measured Moho velocity beneath it.

Moreover, when our velocities from SB 1009, SB 1010 and SB 1011 are plotted on the coincident reflection profile, our 4.8 km/s arrival on SB 1009 and SB 1010 corresponds with a major change in the reflection profile to lower frequency, less parallel arrivals suggesting the top of basement. The velocity of the crust ranges from 4.8-5.3 km/s consistent with continental velocities with no indication of oceanic layer 3. Near SB 1011 the possible basement horizon changes character (Fig. 14). A series of high amplitude, probably sedimentary, reflections are observed infilling an irregular basement structure. The velocity function for SB 1011 is typical of transitional crust with a 7.5 km/s measurement. Similar topographic irregularities in base of the sedimentary layer and the change to serpentinized crust also

occur on the North-South transect (Fig. 12). The magnetic map (Fig. 7) shows a strong positive anomaly in the vicinity of SB 1009 and SB 1010 (Fig. 1) becoming subdued at the location of SB 1011.

5.1.3 Near the Canadian polar margin

Along the northern Canadian margin (Figs. 1 and 10), Berry and Barr (1971) conducted a refraction experiment that had sparse recording and shot intervals. They measured a crustal velocity of 5.4 km/s and a Moho velocity of 8.1 km/s at a poorly resolved depth of 15 ± 9 km mid-way down the continental slope, suggesting thinned continental crust. Two sonobuoys near to, but seaward of the Berry and Barr (1971) refraction line (Fig. 1) provide additional data for the crust in deeper water. SB 1020, slightly south and seaward of the Berry and Barr (1971) line has a well resolved 6.6 km/s layer. The sonobuoy closest to the continental margin, SB 0834, (Fig. 9) recorded data sporadically beyond 17 km offset; however, there were indications of refraction with a 7.4 km/s velocity typical for transitional crust.

In summary, the earlier refraction experiments in the Canada basement can be put in context with our results. Most of these experiments were along the edges of Canada Basin and did not have coincident reflection profiles making it difficult for the extent of the thick sedimentary deposits to be resolved. The high velocities in the range of 4.5-4.9 km/s in sedimentary section were often interpreted as oceanic layer 2. None of the earlier experiments identified a clear oceanic layer 3 velocity. Based on their distribution in the periphery of Canada Basin, none of them were actually located on oceanic crust. Our more closely sampled sedimentary and upper crustal wide angle reflection/refractions, the magnetic patterns and reflection profiles suggest these experiments actually recorded a sedimentary section overlying transitional or thinned continental crust, consistent with their lack of oceanic velocities.

5.2 Constraints on the Opening of Canada Basin

The sonobuoy velocities, seismic reflection profiles and potential field data consistently describe a region of typical and distinctive oceanic crust in Canada Basin (Fig. 15), primarily associated with the northern segment of the CBGL. Our data distribution is insufficient to assign crustal type near and to the west of the southern segment of the CBGL. The paired magnetic anomalies and the general symmetry of oceanic crust around the CBGL, are consistent with the counter clockwise rotation of Alaska with a pole of rotation located in the south of the Beaufort Sea (e.g., Grantz et al., 2011, and included references).

The combined data sets indicate that the area of oceanic crust is a sub-region of the larger Canada Basin (Fig. 15) and a plate reconstruction must also account for the unknown but finite extension in the transitional crust (e.g., Alvey et al., 2008; Müller et al., 2008). The added space from this more restricted closure might accommodate parts of Northwind Ridge and the Chukchi Borderland, which otherwise require special tectonic accommodation to avoid the overlap with the Canadian Arctic Islands (e.g., Grantz et al., 2011; Brumley et al., 2015). Detrital zircon analysis identify that the Alaska Beaufort margin was in close proximity with the opposing margin of the Canadian Arctic Islands during the Ellesmerian Orogeny into Mesozoic time (Gottlieb et al., 2014). The closure proposed in Gottlieb et al. (2014) results in a 55° rotation, or 10° less than the 66° rotation proposed by Grantz et al. (2011). The width of the southern edge of the polygon of oceanic crust (~340 km wide), puts limits on the amount of rotation, assuming that the rotation is symmetrical around a pole south of the Beaufort Sea. However, the new data presented here and onshore constraints (Lane et al., 2015) indicate that the rotational model does not satisfy all the observations.

5.3 Rate and Age of Seafloor Spreading

The thickness and morphology of oceanic crust can augment magnetic anomalies in determining the spreading rate of an ocean basin. Oceanic crustal thicknesses of 4 to 7 km in Canada Basin are typical of

slow- spreading rates, i.e., less than full spreading rates of 20 mm/a that produce 6 km thick oceanic crust at average mantle temperatures (Reid and Jackson, 1981; Bown and White, 1994). Faster spreading rates produce average thicknesses of 7.1 km (White et al., 1992).

The morphology of the axial graben also helps constrain spreading rate (MacDonald, 1982) because slower spreading rates are associated with deeper axial grabens. The depth of the axial graben in Canada Basin based on the seismic reflection data (Fig. 9), is ~1.5 km, consistent with 1.5-3 km depths typical of axial grabens formed by spreading rates of 10-20 mm/a. This depth is greater than the estimate of Grantz et al. (2011), which was based on a single partial crossing of the axial graben. High relief and roughness of the basement surface observed in the multichannel data is also consistent with expected roughness from slow and ultraslow spreading ridges (Malinverno, 1991; Ehlers and Jokat 2009).

Regional geology can help constrain the time of rifting and extension and therefore also places limits on the age of opening. Based on onshore geological constraints such as the dating of onshore rocks assigned to the HALIP, the breakup unconformity, offshore heat flow data and paleomagnetic data on the rotation, seafloor spreading in Canada Basin was estimated to have initiated in the Hauterivian (~134 Ma, Embry and Osadetz, 1988; Embry and Dixon, 1994; Estrada, 2015) and slightly younger offshore of Alaska (Hubbard et al., 1987). The width of the oceanic crust at its southern limit is about 320 km from the outside of the paired magnetic anomalies. The outer magnetic anomaly is more continuous than the inner that is formed by a series of small circular highs. This could be due to the magma generation processes slowing down as the spreading center became extinct. Our sonobuoy velocities indicate the crust near the axial valley does not have all the crustal layers observed elsewhere. A slow spreading rate of 20 mm/a would require about 15 Ma to produce this extent of crust. An

818 ultraslow rate of less than 20 mm/a is not consistent with the development of layer 3 velocities (Jokat et
819 al., 2012).

820
821 To correlate the seafloor magnetic anomalies in Canada Basin with the magnetic reversal sequence two
822 linear positive magnetic anomalies with an intervening negative anomaly (Fig. 7) are identified. This
823 reversal sequence cannot happen between ~124 and ~83 Ma due to the long Cretaceous positive
824 polarity chron. Likewise, prior to M5 (~130 Ma), the many short reversals are unlikely to be preserved in
825 a slow spreading ocean basin. Both positive and negative reversals occur in the intervening time interval
826 of ~130 to ~124 Ma (i.e., between M5 and M0). This age is consistent with the Early Cretaceous age of
827 opening based on geological considerations (e.g., Embry and Dixon, 1994).

828
829 In order to fit the width of the negative anomaly (i.e. the trough between the two positives) with an
830 assumed spreading rate of 20 mm/a, 5 Ma of time is required. However, the longest negative anomaly
831 duration between M5 and M0 is M3, about 2.3 Ma which would require a higher spreading rate of ~35
832 mm/a to fit the anomaly. This leaves us with an unresolved problem, that a higher spreading rate (e.g.,
833 of 35 mm/a) is required to create the width of the oceanic crust or the time frame is incorrect. If the
834 spreading rate is high, the thickness of oceanic crust should be consistently greater than the observed 4-
835 7 km. A similar identification of magnetic chrons of M4n to M2n (130-127.5 Ma) has been suggested by
836 Grantz et al. (2011). This even more limited age range requires a spreading rate of 75 mm/a that is also
837 inconsistent with anomalously thin oceanic crust measured from sonobuoy velocities. Another
838 possibility is that the paired magnetic anomalies are younger. A single whole rock sample from Alpha
839 Ridge was dated using $\text{Ar}^{40}/\text{Ar}^{39}$ incremental heating with a plateau age of 89 ± 1 Ma (Jokat et al., 2013).
840 The reversal sequence at the end of the long positive chron C34 would be a better fit for a slow
841 spreading ocean. A third possibility is that the magnetic anomalies are due to basement topography.

5.4 Seafloor Spreading

The sonobuoy velocity data do not uniquely constrain whether seafloor spreading preceded, coincided with, or followed the magmatism associated with the HALIP. The inferred age of Canada Basin spreading between ~130-~124 Ma closely coincides with onset of magmatism in the HALIP, 130 Ma (Maher, 2001), and revised duration of magmatism in the HALIP of 138-125 Ma (Døssing et al., 2013) or possibly 130-101 Ma (Evenchick et al., 2015). The sparse dates available from onshore dykes that radiate from Alpha Ridge, the offshore segment of the HALIP, suggest basaltic magmatism starting at about 124 Ma (Bleeker et al., 2012; Corfu et al., 2013). The age of bedrock sample from Alpha Ridge 89 ± 1 Ma (Jokat et al., 2013) suggests magmatism on the Alpha Ridge may have had a broad time span. The earliest dates suggest that the creation of oceanic crust in Canada Basin for the region encompassing the northern segment of the CBGL was coeval with initial formation of Alpha Ridge. This timing is probably also the same as formation of the volcanic continental margin at the junction of Alpha Ridge with the Canadian Arctic Islands margin (Funck et al., 2011).

The age of opening of southern Canada Basin, along the southern segment of the CBGL where it projects ashore near the Mackenzie Delta (Fig. 15), is not well constrained. Transitional crust has not been sampled or dated and is buried by up to 12 km of sedimentary strata beneath the Mackenzie Delta/Fan depositional system (Shimeld et al., 2016). Southern Canada Basin has several characteristics that distinguish it as a separate crustal entity: the more subdued magnetic character, the different orientation of the CBGL relative to the region of oceanic crust further north, the consistently higher sedimentary velocities (Shimeld et al., 2016) and the inferred existence of the oldest sedimentary units along the Beaufort margins (e.g., Hutchinson et al., 2012; Mosher et al., 2013). Extensional basins may have begun forming as early as the Early Jurassic (Hubbard et al., 1987; Embry, 1991; Grantz et al., 1994;

Harrison and Brent, 2005). Whether the transitional crust along the southern segment of the CBGL represents an early (i.e., Jurassic or Early Cretaceous) opening or extension, or is coeval with the formation of oceanic crust requires additional study and modeling.

5.5 The nature of the ocean-continent transition

The border drawn around the ocean crust polygon (Figs. 7, 8, and 15) defines the ocean-continent boundary (OCB), i.e., the location where the seaward edge of transitional crust and the landward edge of oceanic crust meet. The OCB is the most controlled crustal boundary based on our data. It corresponds to the edge of the ocean crust based on the velocity classification (Fig. 3), the edge of the paired linear magnetic anomalies (Fig. 7), and the transition from irregular basement highs on oceanic crust to smoother and, on the eastern, Canadian side, deeper basement on transitional crust on the seismic reflection records (Figs. 9-13).

Outside of the OCB, the seismic data distribution is more widely spaced leading to more uncertainty in characterizing the heterogeneous nature of the ocean continent transition (OCT), i.e., the zone where rapid changes in crustal thickness, lithology, and deformation associated with rifting occurs (Fig. 15). The landward extent of the OCT can be inferred from the integration of the seismic velocity, reflection character, seafloor morphology and potential field anomalies (solid and dashed lines in Fig. 15), but is largely unconstrained from lack of data under the thicker sedimentary cover at the intersection of the Chukchi Borderland with Alaskan margin and along the Canadian Beaufort margins (question marks in Fig. 15a). Nor have we attempted to map transitional crust towards the HALIP as it was mapped by Saltus et al. (2011, white dashed line, Fig. 15). Because of the short hydrophone array necessitated for towing in ice, ice conditions that prevented reaching some parts of the margins, and limits in data collection to water depths generally deeper than 2000 m for cruise objectives, our data are not

positioned to give full crustal transects from continental to oceanic crust. Nor are data sufficient to subdivide the OCT into proximal, distal, necking, coupled, exhumed, or inner and outer zones as defined by other authors (e.g., Peron-Pinvidic et al., 2013; Karner et al., 2012; Sutra and Manatschal, 2012).

The lower crustal velocities used to define the OCT show trends around the OCB. The OCT around the northern extent of oceanic crust has velocities of 7.2-7.3 km/s (described in section 4.2.2). The OCT around the southern extent of oceanic crust has consistently slightly higher velocities of 7.4-7.5 km/s. The only sonobuoy record in southern Canada Basin with a velocity of 7.3 km/s (SB 0709) is very noisy with the direct arrival extrapolated from a weakly constrained wide angle reflection, rather than being based on an observable refraction. The remaining 25 velocity classifications of transitional crust are uniformly 7.4-7.5 km/s. Although these differences are near the limits of resolution of the refraction measurements, their consistency across a large data set suggests the trend of slightly lower velocities in the north and slightly higher velocities in the south is real.

There are good geological explanations for the velocity trends. The 7.2-7.3 km/s slightly lower velocities are consistent with expected velocities from gabbroic melts (e.g. Lundin and Dore, 2011) and highly intruded continental crust such as found off Norway (Mjelde et al., 2005), Hatton-Rockall (Funck et al., 2008), the Canadian volcanic margin at Alpha Ridge (Funck et al., 2011), and southeastern Brazil (Evain et al., 2015). The OCT around the northern extent of oceanic crust is likely to be a magma-rich OCT that is dominated by magmatic intrusion and/or underplating in the lower crust.

The higher velocities of 7.4-7.5 km/s are consistent with velocities interpreted to be serpentinized mantle measured and sampled off the southern Iberia (Whitmarsh et al., 1993; Chian et al., 1999; Dean et al., 2000; Clark et al., 2007). Exhumed and serpentinized continental mantle is one of the dominant

characteristics of magma-poor rifting and continental breakup such as documented off Iberia (Whitmarsh et al., 2001; Lavier and Manatschal, 2006; Reston, 2009; Lundin and Dore, 2011). The OCT in this southern area is likely to be representative of a magma-poor OCT that is dominated by serpentinization of exhumed continental mantle.

A few seismic refraction and reflection criteria exist to help resolve between magma-rich and magma poor settings, and our seismic data locally contain additional information. In a magma-poor setting, one would expect a weak to non-existent Moho arrival because of the gradual change from serpentinized to normal mantle over some depth range (e.g., Chian et al., 1999). In magma-rich setting, the Moho is often a strong and distinct arrival (e.g., White et al., 2008). The offsets used in Canada Basin sonobuoy data are not generally large enough to get robust Pn (Moho) arrivals, but one strong Pn arrival off the Chukchi Borderland (SB 1013, star in Fig. 15) which yields a crustal thickness of 5 km is another supportive piece of evidence that the OCT in that region is magma-rich. This thickness is within the range that might be expected with underplated or intruded magmatic material (Blaich et al., 2011). This Pn measurement and its association with a positive magnetic anomaly that appears to be continuous with Alpha Ridge magnetic domain of Saltus et al. (2011, Fig. 15b) supports the origin of this crust to be a mafic-rich, or underplated/intruded body.

The crust above the high-velocity lower crust in magma-rich continental margins can have a complex distribution of basalt flows (seaward dipping reflections), sedimentary strata, and variable velocities as well as variable thicknesses along transects (Peron-Pinvidic et al., 2013; Blaich et al., 2011 and references therein). In our Canada Basin data, there are weak seaward dipping reflections in the north (SDR in Fig. 9b), consistent with the magma-rich tectonic setting. Unfortunately, in the eastern and southern (Beaufort) regions, the top of basement is often not imaged by the reflection data because of

interference from the water-bottom multiple. Hence the evidence for possible magma-rich features such as seaward dipping reflections in these regions is obscured. Similarly, the small airgun source, the short streamer offsets, and the resulting strong water-bottom multiple also prevent imaging of a possible mid-crustal detachment surface that is often characteristic of magma-poor continental margins (Reston, 2009).

Moho is also inferred in our dataset from PmP (or reflections from Moho) or occasionally Pn (mantle refractions), in approximately one quarter of the sonobuoys in the OCT. These PmP reflections have strong or intermittent reverberatory arrivals suggesting a relatively sharp, rather than gradational, interface at the top of upper mantle. PmP arrivals are often strong second arrivals at offsets less than 30 km and are therefore not affected by degraded signal to noise at more distant offsets. In Canada Basin sonobuoys, the PmP arrivals are distributed around the north, west, and south of oceanic crust. Their distribution about the zone of oceanic crust, similar to the sonobuoys that lack PmP arrivals, suggests that the upper mantle Moho is heterogeneous at the scale of this seismic experiment. This observation does not necessarily eliminate serpentinized upper mantle as an interpretation because the actual velocity measurements of 7.4-7.5 km/s are still diagnostic. However, the number and distribution of PmP arrivals suggest more complexity than the simple gradational velocity model provides. Lacking shear waves, our sonobuoy data cannot be used to calculate V_p/V_s ratios that could perhaps resolve differences between intruded crust and serpentinized mantle (Mjelde et al., 2002).

A final indicator of the heterogeneous nature of the OCT is its variable width (Fig. 15), which ranges from 20 km or less immediately north of Alaska to ~250 km in the vicinity of the Mackenzie delta/fan. The width is ~200 km along the central Chuckchi Borderland and tapers northward towards the HALIP to less than 100 km. It is well known the widths of the OCT on conjugate margins of the North and South

Atlantic are asymmetric and variable, and can be narrower than ~50 km or wider than ~200 km (Peron-Pinvidic et al., 2013; Blaich et al., 2011 and references therein). The widths measured around Canada Basin are therefore consistent with measurements and models from other passive margins.

Two blocks of crust interpreted to have continental crustal velocities abut the two narrowest zones of the OCT, one in the south (Fig. 15; labelled and discussed earlier to coincide with the D3 magnetic domain of Saltus et al., 2011) and one in the north (labelled 78 N, Fig. 15a). Narrow OCTs are often associated with transform margins (Scrutton, 1979), but such an association is not compelling for these regions in Canada Basin. Transform faults are not well developed at ultra-slow spreading ridges such as the Knipovich or Gakkel Ridges in the Arctic) and the magnetic data in Canada Basin do not show offsets that could be interpreted as transforms, either at the locations of these narrow OCTs or elsewhere (Fig. 15c). The basement depression at 78°N (Fig. 15) is a steep-sided sedimentary filled basement deep (Fig. 9) 80 km by 200 km in extent that is observed in the seismic and potential field data. More detailed investigations to determine the extensional or strike/slip origin would provide important constraints in understanding the development of Canada Basin.

6 Conclusions

The most significant result, of the velocity analyses of the broadly geographically distributed sonobuoy data, is that only the central part of Canada Basin is underlain by velocities typical of oceanic crust. When the results of the assignments of velocities to crustal type are superimposed on the magnetic map, we see only the regions with oceanic layer 3 velocities have linear anomalies typical of seafloor spreading. In addition, the CBGL that has been previously attributed to the fossil spreading center bisects the region identified as oceanic crust. Examination of the coincident seismic reflection profiles

shows a deep sediment filled valley that is located along the northern segment of the CBGL. The seismic reflection profiles with coincident oceanic crustal velocities show that the top of seismic basement has high-relief crust and low reflectivity. Velocities determined to be typical of transitional crust are observed on the periphery of the oceanic crust. On the reflection profiles the top of the transitional crust has strong reflectivity and little relief. The magnetic data in the transitional crustal zone have amorphous shaped anomalies. The junction of the oceanic and transitional crust, based primarily on seismic refraction velocities and magnetic character supported by the seismic reflection profiles, is the OCB.

In the Beaufort Sea and further north along the Canadian Arctic Islands (south of the CBGL), velocities typical of serpentinized mantle (7.4-7.5 km/s) are observed in Canada Basin. A serpentinized zone near the margins is a typical feature of slow or oblique spreading oceans. The slow spreading is consistent with the thin oceanic crust (4-7 km thick) and the deep central axis of the fossil spreading center (up to 2 km). In contrast, on the northwestern edge of Canada Basin, seaward of Northwind Ridge in the vicinity of a high-amplitude magnetic anomaly NoRMA, there are several sonobuoys that have velocities in the 7.2-7.3 km/s range. These sonobuoys are also located near the north end of the oceanic zone and close to a region dominated by high amplitude anomalies associated with Alpha Ridge. On this side of Canada Basin, the transitional velocities are interpreted to be typical of a magma rich margin.

Regions of thinned continental crustal velocities of variable thickness are inferred around the edges of Canada Basin. Where identified offshore Alaska and in the northern Canada basin, thinned continental crust occurs in water depths of 3500 m or greater, indicating it reaches far offshore of the continental slope. Published long offset refraction profiles, did not record oceanic layer 3 velocities supporting our interpretation. In the southeastern part of the basin, continental crust extends minimally offshore. The

1009 deep crust at^oN basin with a single continental crustal velocity is a distinctive feature on the gravity map.
 1010 This isolated feature with a significant sedimentary section needs further investigation.
 1011 Offshore of the Alaska margin, a broad region with velocities and thickness appropriate for thinned
 1012 continental crust is predicted from the velocity determinations of the sonobuoy data. Furthermore, the
 1013 magnetic signature in this region has longer wavelengths than observed elsewhere in Canada Basin and
 1014 may be an offshore continuation of the broad high-amplitude magnetic anomaly domain (D3) that
 1015 onshore is associated with thick continental crust, high mafic crustal content, and possibly strong
 1016 lithosphere.
 1017
 1018 The most popular hypothesis for the opening of Canada Basin is a counterclockwise rotation of Alaska
 1019 away from the Canadian Polar margin. The magnetic anomalies in the limited oceanic crust area are
 1020 consistent with this model. However, although oceanic crust is nearly symmetrically about the CBGL, the
 1021 asymmetrical distribution of the oceanic crust with respect to the margins indicates that a more
 1022 complicated plate tectonic model is needed to explain the observed data.

1023

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1035

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9 Figure Captions

Fig. 1. a) Location of data in Canada Basin. Oceanic, transitional and continental crustal velocities from analysis of the sonobuoys are marked with white, green and black symbols, respectively. Thin solid black lines: coincident seismic reflection profiles. Yellow/cyan lines: assembled seismic transects shown in the paper. Dotted yellow line: Canada Basin Gravity Low (CBGL). Dotted red lines indicate basement depression. Also shown are previous crustal scale profiles: near Northwind Ridge (Jackson et al., 1995), near Alaskan margin (Baggeroer & Falconer, 1982; Mair & Lyons, 1981), and across Canadian Arctic margin (Stephenson et al., 1994 and Berry & Barr, 1971). b) Enlargement with sonobuoys labeled. Small labelled grey boxes indicate sonobuoys with undefined crustal type.

Fig. 2. Idealized transects of (a) magma-rich (volcanic) passive margin showing the seaward dipping reflections (SDRs), and intruded/underplated body, modified from Blaich et al. (2011); and (b) magma-poor (non-volcanic) margin showing the exhumed continental mantle adjacent to oceanic crust, modified from Mohn et al. (2012). OCB: oceanic-continent boundary, OCT: ocean-continent transition zone. Velocities shown are values used in this paper.

Fig. 3. Observed velocity/depth curves, excluding the sedimentary layers, determined from analyses of the wide angle reflection/refraction data (after Chian & Lebedeva-Ivanova, 2015) are plotted in three groups: a) continental, b) oceanic and c) transitional. The grey zone illustrates the range of velocities compiled for oceanic crust at age of 142-143 Ma in North Atlantic Ocean (after White et. al., 1992). The groupings are based on the velocity criteria shown in Table I. Second row of boxes show the same curves as the first row but colored and sized for easier labelling.

Fig. 4. Sonobuoy 0726 as an example with oceanic crustal velocities (for location see 726 in red in Fig. 1).

a) The reflection profile along the model. b) Same shots recorded by the sonobuoy, modelled and plotted by Reduced Normal Move-Out (NMO) transform such that each wide-angle reflection phase (dotted blue) at zero offset matches exactly with the corresponding event on the reflection profile. Sedimentary velocities were modelled by Chian & Lebedeva-Ivanova (2015) and used for modelling deeper arrivals with labels indicating their corresponding velocities. Inset of b) shows crustal refraction phases picked (yellow bars) for statistical error analysis with results shown to its left; every trace is picked. c) 1D velocity/depth curve at the sonobuoy location. d) Final 2D velocity model from the forward modeling, with 5 sets of ray paths illustrated. Each model layer is parameterized by a linear vertical velocity gradient with no horizontal changes in velocity inside the same layer.

Fig. 5. Sonobuoy 1034 as an example with transitional crustal velocities. See 1034 (in red) in Fig. 1 for location and Fig. 4 for caption.

Fig. 6. Sonobuoy 1003 as an example with continental crustal velocities. See 1003 (in red) in Fig. 1 for location and Fig. 4 for caption.

Fig. 7. Magnetic anomaly map of Canada Basin (after Gaina et al., 2011). Sonobuoy models classified as oceanic are shown as white squares, continental as black squares and transitional as blue diamonds. Dashed polygon indicates ocean-continent boundary (OCB; Table 2). The number of each solid blue line indicates the figure number of the seismic transects. D3 indicates a regional magnetic zone defined by Saltus et al. (2011). NoRMA: Northwind Ridge Magnetic Anomaly. Parallel white dots indicate a basement depression.

Fig. 8. Free Air gravity anomaly map of Canada Basin (with onshore Bouguer anomaly; after Andersen et al, 2010). Sonobuoy models classified as oceanic are shown as white squares, continental as black squares and transitional as blue diamonds. Dashed polygon indicates the ocean-continent boundary (OCB; Table 2). The number of each solid blue line indicates the figure number of the seismic transects. Parallel red dots indicate a basement depression.

Fig. 9. Seismic transect across the northernmost Canada Basin (see Fig. 1 for location). a) Observed magnetic and gravity anomalies (extracted from gridded maps in Figs. 7 and 8; CBGL: Canada Basin gravity low. C|C indicates course change. b) Reflection profile (Hilbert transformed) converted to depth using a velocity model based on all available sonobuoy data along the transect. The time-depth conversion process includes: assign distance to each trace of reflection profile, construct a velocity model for sediments between seafloor and basement in time domain, and perform time-depth conversion; final crustal model is constructed in depth domain that in general matches all available data. Sonobuoy models in the crust along the transect are overlain as solid lines with labels showing their velocities in km/s; brackets indicate offline stations. HVLC: High velocity lower crust. SDR: Seaward dipping reflectors. The rectangular area marked on panel b) is enlarged in c) to show details (vertical exaggeration: 7.6:1). Vertical arrows indicate base of sedimentary layers.

Fig. 10. Seismic transect across the central Canada Basin (see Fig. 1 for location). Top: magnetic (blue line) and gravity (red line) anomalies extracted from gridded data in Figs. 7 and 8, with computed Free Air gravity anomaly (black line) for Moho constraint. Gravity computation used a velocity-density curve from Ludwig et al. (1970). Computed gravity is DC-shifted by -712 mGal and assumes a constant density of 3.3 g/cm^3 underneath the model. Bottom: velocity model

overlain by depth-converted reflection profile. Sonobuoy 1021 is indicated in blue to contrast with adjacent instruments. See Fig. 9b for more description.

Fig. 11. Seismic transect across southern Canada Basin (see Fig. 1 for location). a) Magnetic (blue line) and gravity (red line) anomalies extracted from gridded data in Figs. 7 and 8. C|C indicates course changes. b) Velocity model overlain by depth-converted reflection profile. Sonobuoy 804 is indicated in blue to contrast with adjacent instruments. The rectangular area marked on panel b) is enlarged in c) to show details of deeper reflectors coincident with 7.4 km/s layer (vertical exaggeration: 11:1). See Fig. 9b for more description.

Fig. 12. Seismic transect from Alaskan margin northward to nearly parallel to Canada Basin Gravity Low (CBGL). Top: magnetic (blue line) and gravity (red line) anomalies extracted from gridded data in Figs. 7 and 8. Bottom: the transect combines 4 seismic profiles, crosses continental, transitional and oceanic crust, and shows distinctive features in crustal velocity structures as well as basement morphology. Reflection profiles are brute stack data converted to depth, with positive amplitudes plotted as grey and negative amplitudes as blue. See figs. 9a and 9b for more description.

Fig. 13. Seismic reflection profile that overlaps the refraction lines reported in Mair & Lyons (1981). The velocities reported by them are plotted in red and the velocities based on our sonobuoy data are in black. Sonobuoy 1002 is categorized as in transitional zone due to an observation of HVLC. But this region is in general continental because: crustal velocity are in the continental range; crust shows consistent reflection characters with regional rotated fault blocks; and it is surrounded by typical continental crust in the vicinity of the shelf.

1386 Fig. 14. Seismic reflection profile that overlaps the refraction lines reported in Baggeroer & Falconer
1387 (1982). The velocities reported by them are shown in red and labeled CANBARX 78.

1388 Fig. 15. Interpreted distribution of crustal regions plotted on a) bathymetric, b) gravity and c) magnetic
1389 maps (for colour bars of scale see Figs. 1, 8 and 7). The regions are based on the combination of
1390 sonobuoy velocities, magnetic and gravity data, seismic reflection control and geological
1391 context. The area with gray shading is the ocean-continent transition. Its inner limit is the ocean-
1392 continent boundary. The star shows the position of sonobuoy 1013. The red and yellow dots
1393 indicate the position of the Baggeroer & Falconer (1982) and Jackson et al. (1995) refraction
1394 experiments. Abbreviations: CBGL-N is Canada Basin Gravity Low-North, CBGL-S is Canada Basin
1395 Gravity Low-South, MR is Magma Rich, MP is Magma Poor, HALIP is High Arctic Large Igneous
1396 Province.

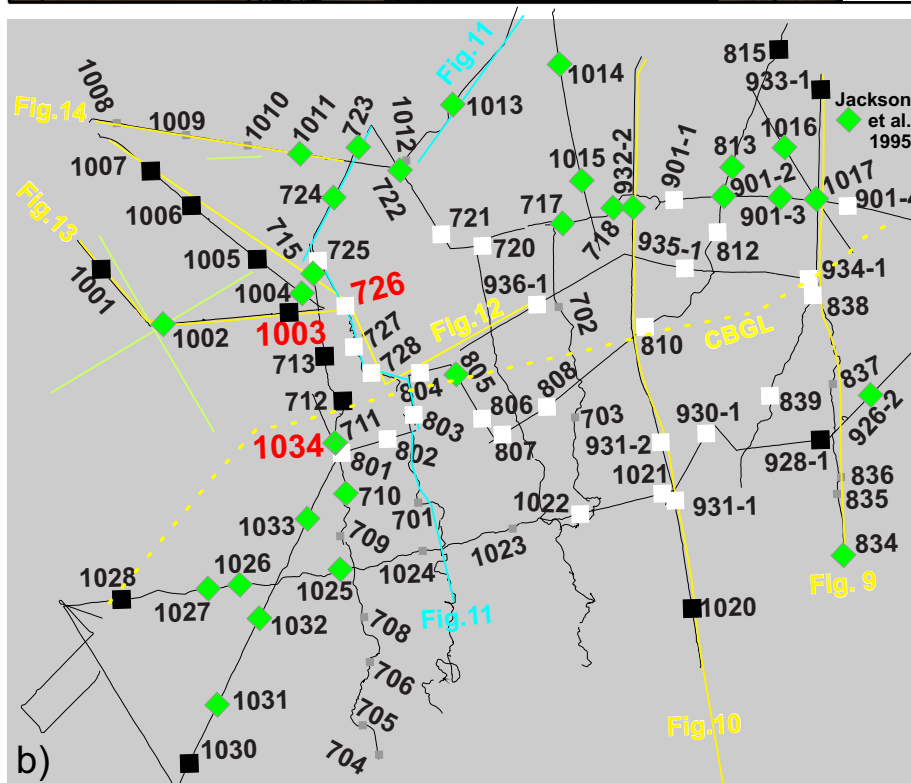
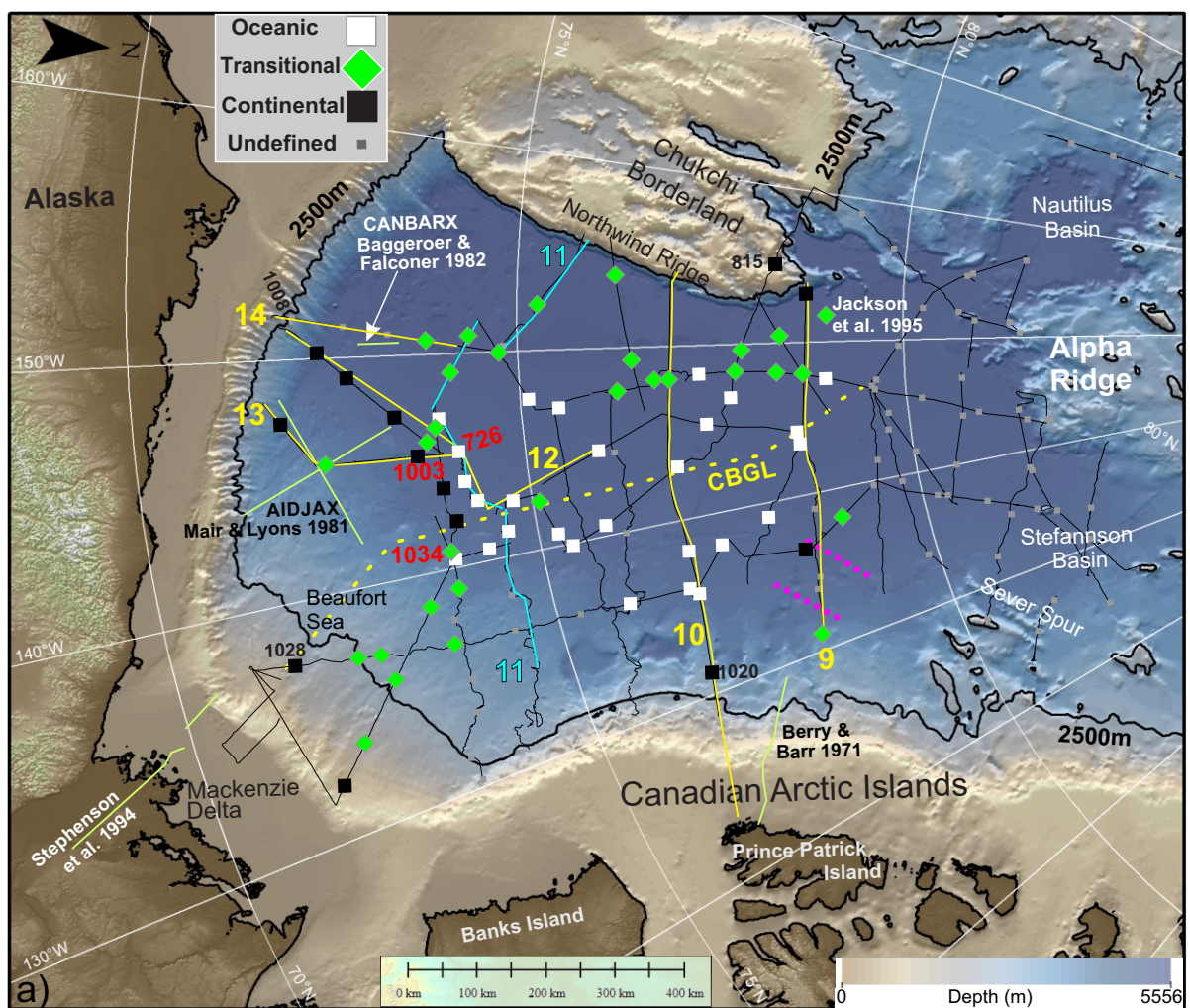


Fig. 1. a) Location of data in Canada Basin. Oceanic, transitional and continental crustal velocities from analysis of the sonobuoys are marked with white, green and black symbols, respectively. Thin solid black lines: coincident seismic reflection profiles. Yellow/cyan lines: assembled seismic transects shown in the paper. Dotted yellow line: Canada Basin Gravity Low (CBGL). Dotted red lines indicate basement depression. Also shown are previous crustal scale profiles: near Northwind Ridge (Jackson et al., 1995), near Alaskan margin (Baggeroer & Falconer, 1982; Mair & Lyons, 1981), and across Canadian Arctic margin (Stephenson et al., 1994 and Berry & Barr, 1971). b) Enlargement with sonobuoys labeled. Small labelled grey boxes indicate sonobuoys with undefined crustal type.

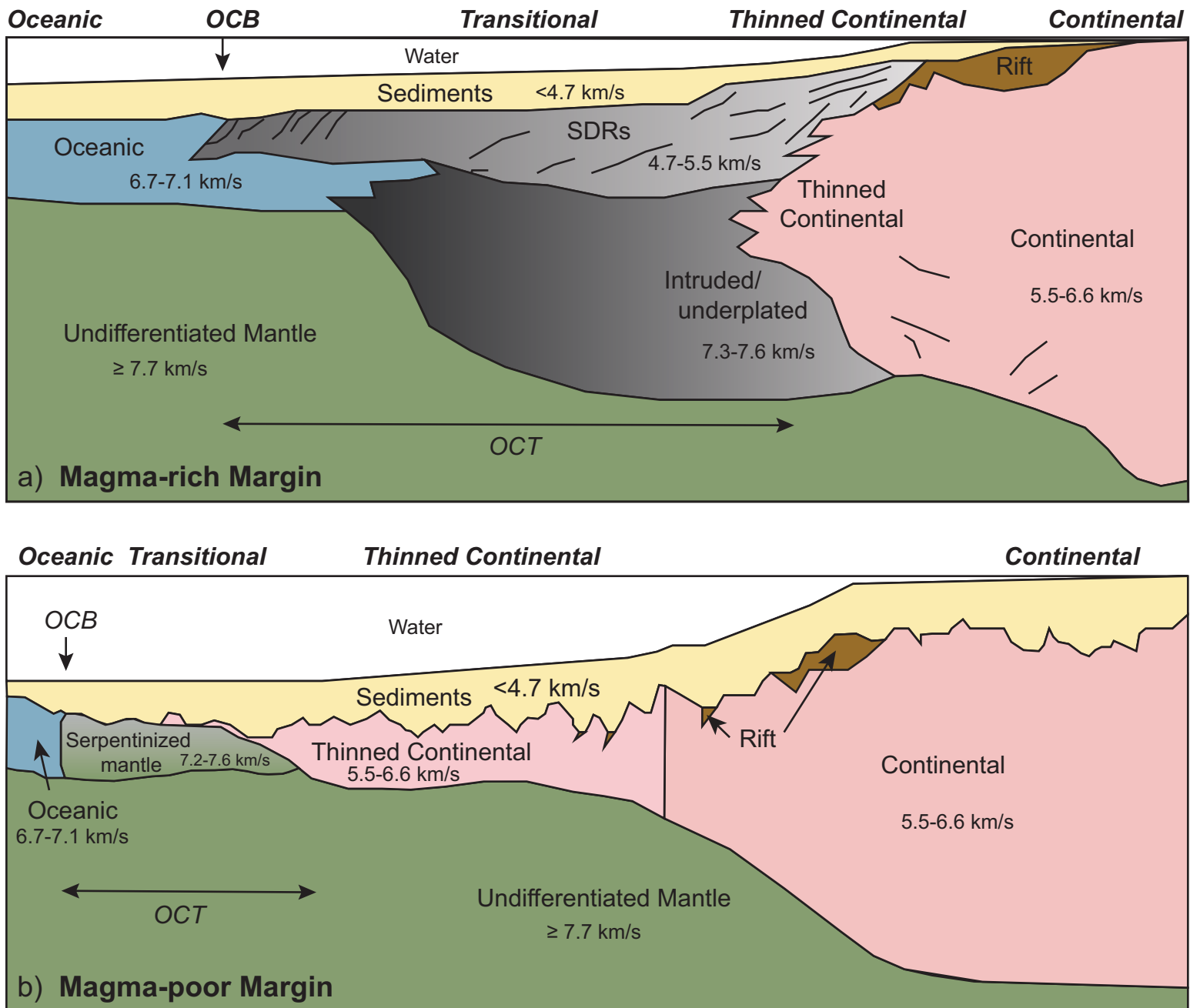


Fig. 2. Idealized transects of (a) magma-rich (volcanic) passive margin showing the seaward dipping reflections (SDRs), and intruded/underplated body, modified from Blaiç et al. (2011); and (b) magma-poor (non-volcanic) margin showing the exhumed continental mantle adjacent to oceanic crust, modified from Mohn et al. (2012). OCB: oceanic-continent boundary, OCT: ocean-continent transition zone. Velocities shown are values used in this paper.

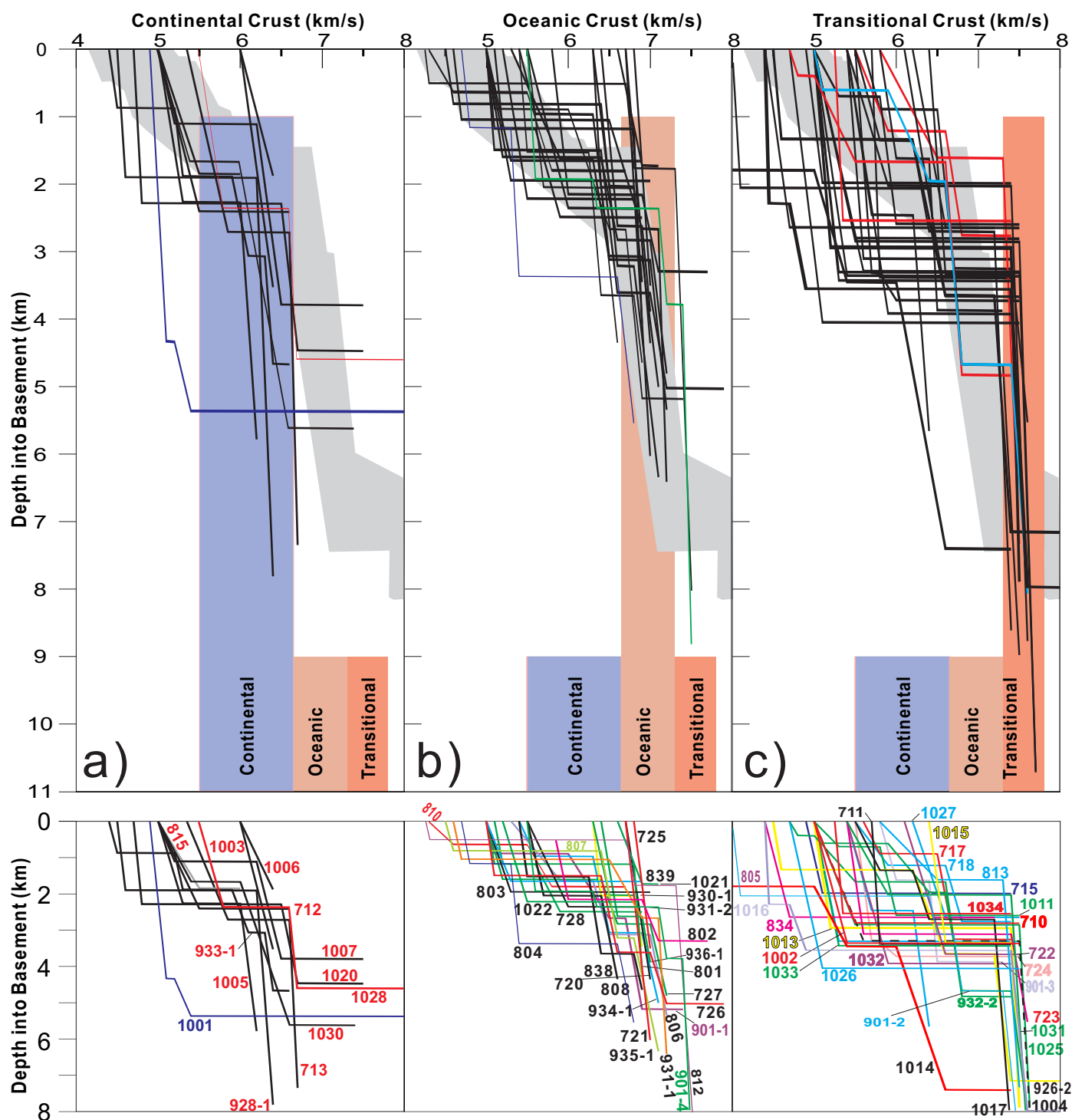


Fig. 3. Observed velocity/depth curves, excluding the sedimentary layers, determined from analyses of the wide angle reflection/refraction data (after Chian & Lebedeva-Ivanova, 2015) are plotted in three groups: a) continental, b) oceanic and c) transitional. The grey zone illustrates the range of velocities compiled for oceanic crust at age of 142-143 Ma in North Atlantic Ocean (after White et. al., 1992). The groupings are based on the velocity criteria shown in Table I. Second row of boxes show the same curves as the first row but colored and sized for easier labelling.

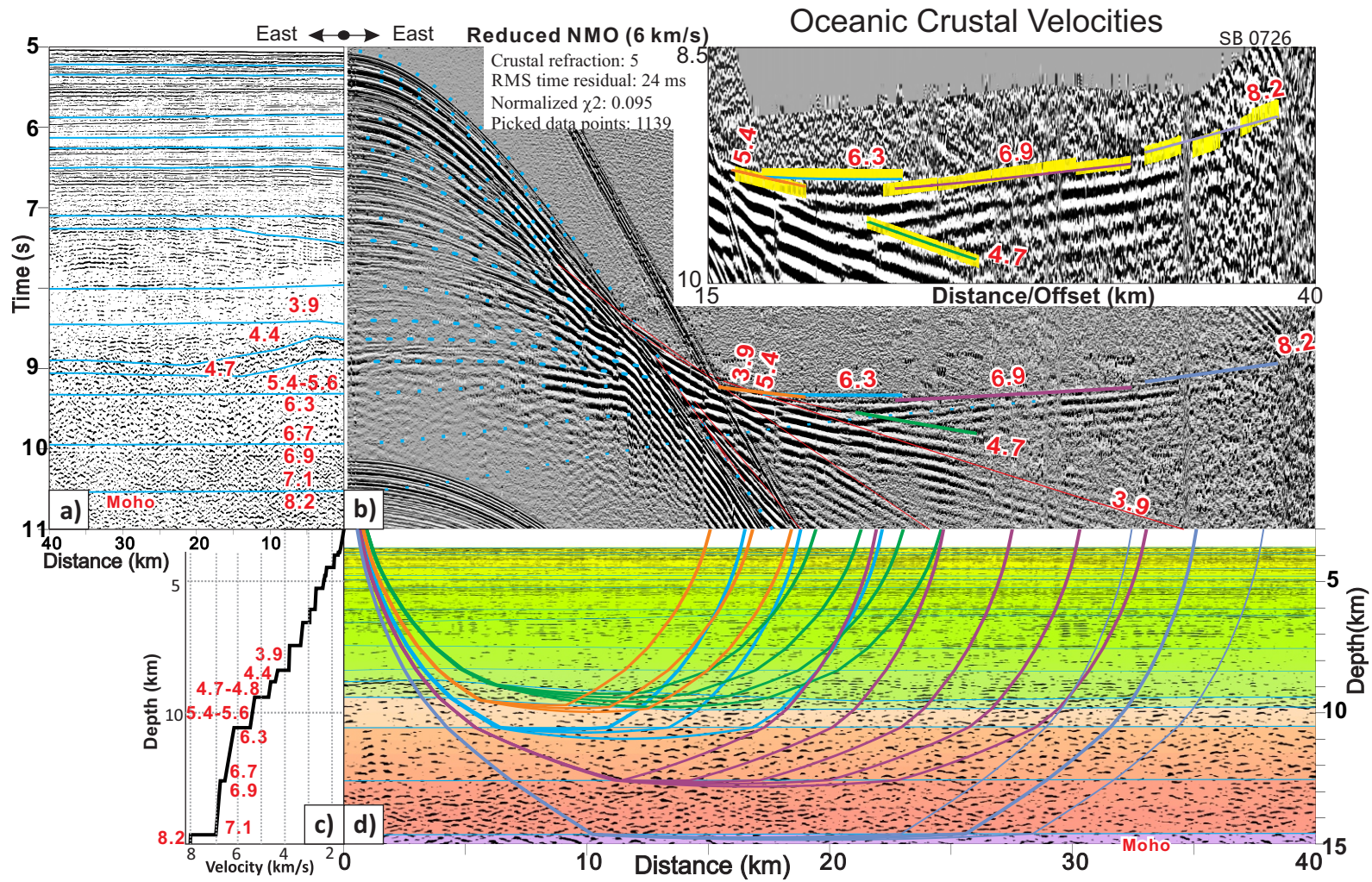


Fig. 4. Sonobuoy 0726 as an example with oceanic crustal velocities (for location see 726 in red in Fig. 1). a) The reflection profile along the model. b) Same shots recorded by the sonobuoy, modelled and plotted by Reduced Normal Move-Out (NMO) transform such that each wide-angle reflection phase (dotted blue) at zero offset matches exactly with the corresponding event on the reflection profile. Sedimentary velocities were modelled by Chian & Lebedeva-Ivanova (2015) and used for modelling deeper arrivals with labels indicating their corresponding velocities. Inset of b) shows crustal refraction phases picked (yellow bars) for statistical error analysis with results shown to its left; every trace is picked. c) 1D velocity/depth curve at the sonobuoy location. d) Final 2D velocity model from the forward modeling, with 5 sets of ray paths illustrated. Each model layer is parameterized by a linear vertical velocity gradient with no horizontal changes in velocity inside the same layer.

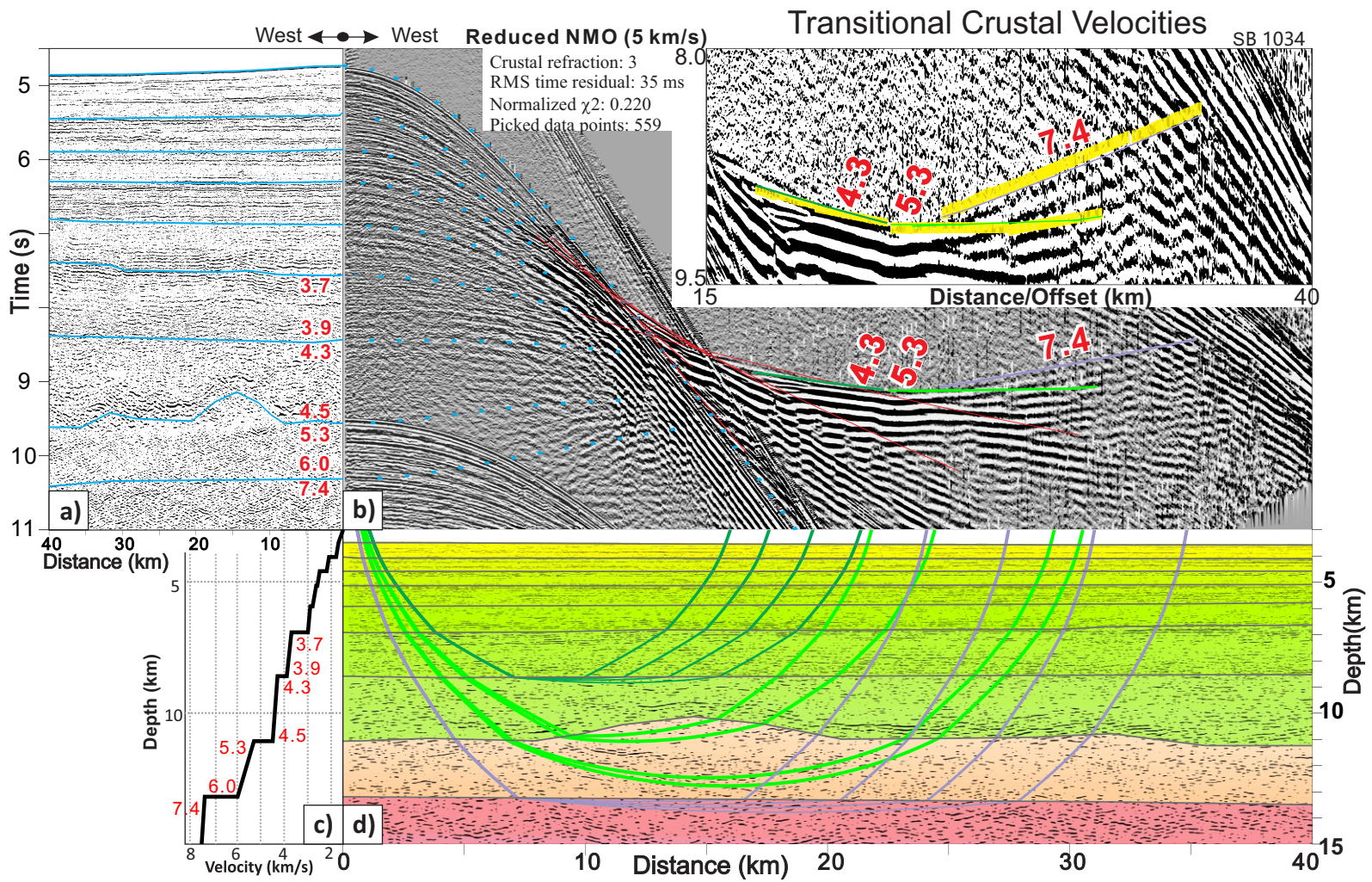


Fig. 5. Sonobuoy 1034 as an example with transitional crustal velocities. See 1034 (in red) in Fig. 1 for location and Fig. 4 for caption.

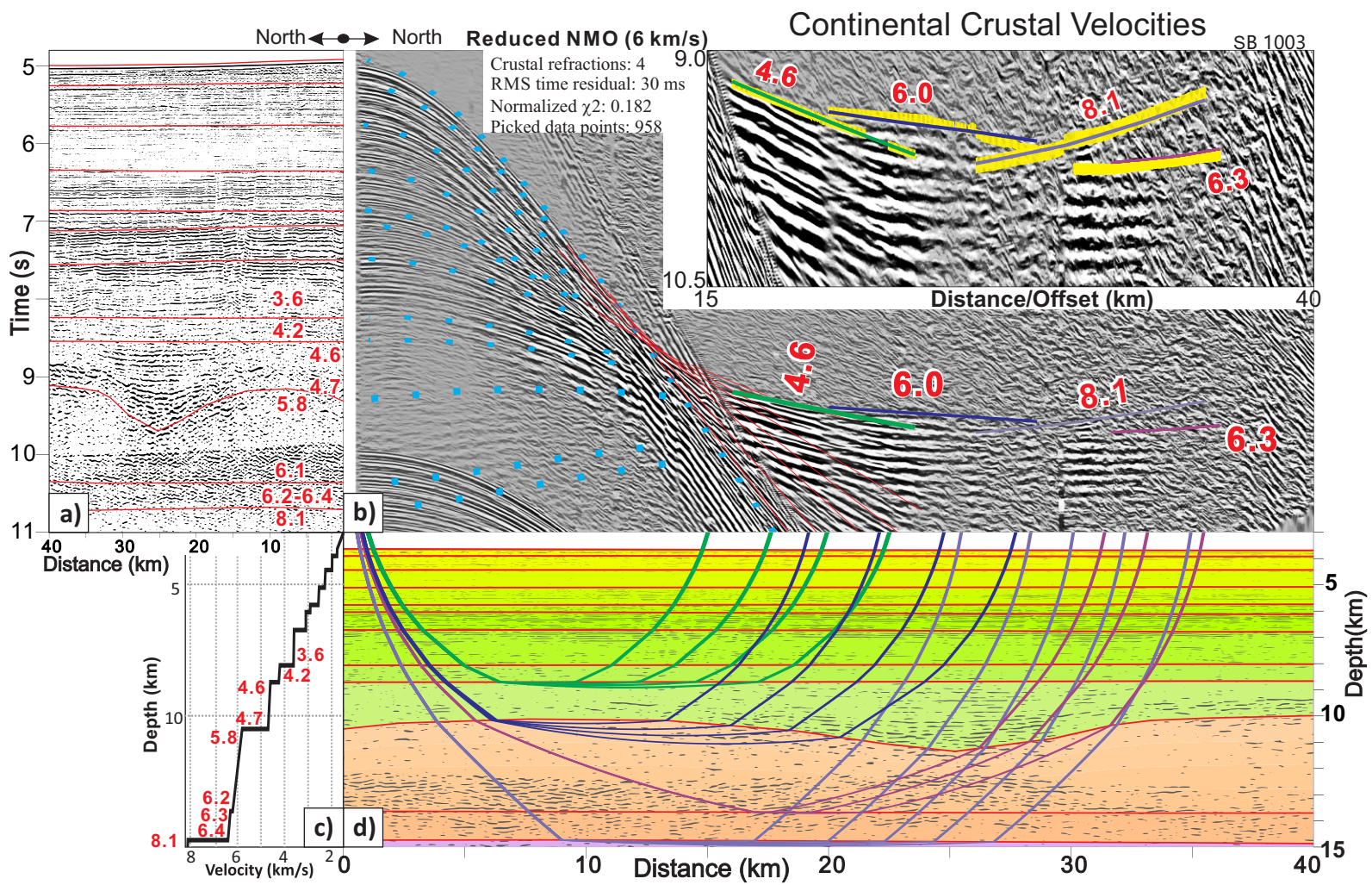


Fig. 6. Sonobuoy 1003 as an example with continental crustal velocities. See 1003 (in red) in Fig. 1 for location and Fig. 4 for caption.

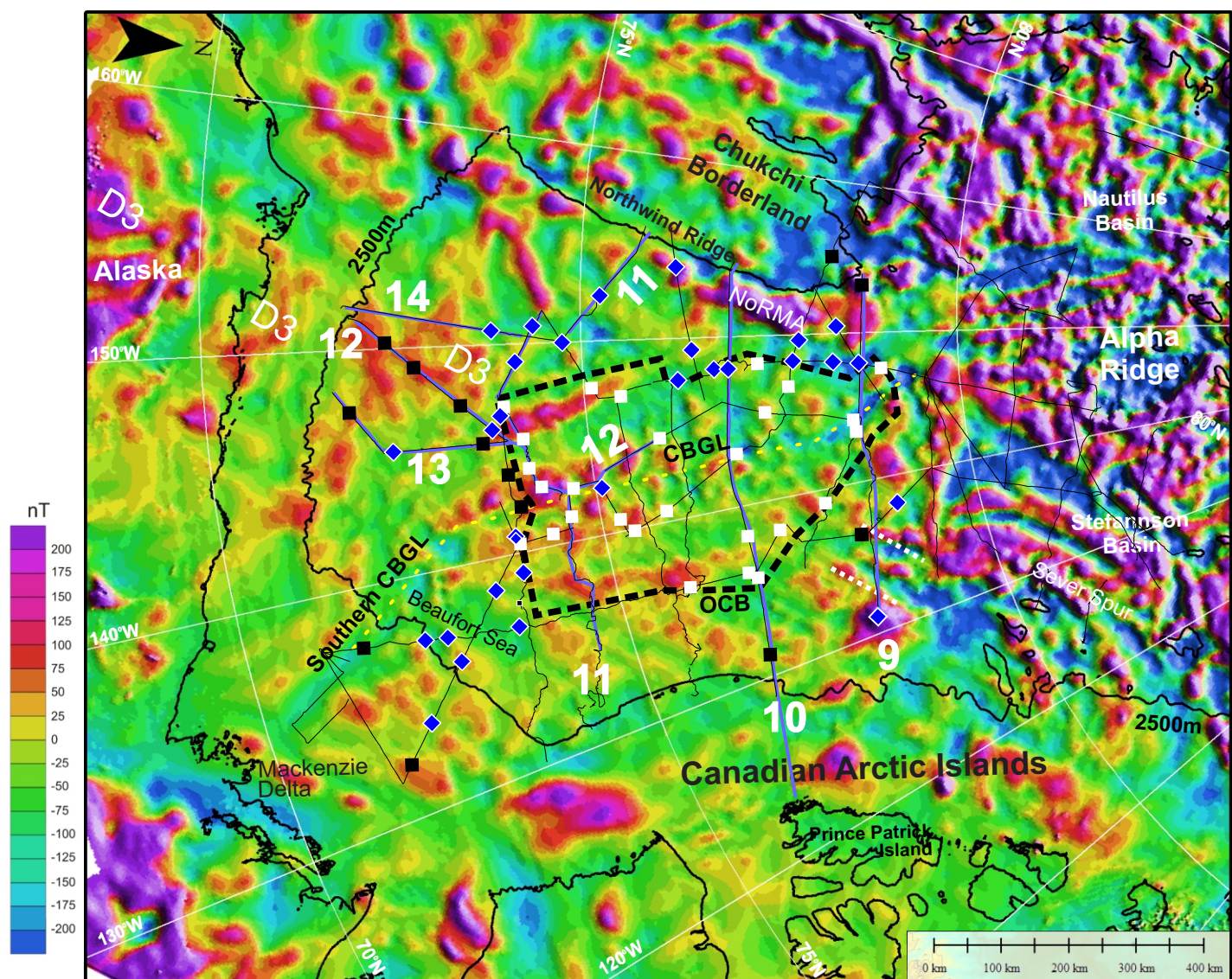


Fig. 7. Magnetic anomaly map of Canada Basin (after Gaina et al., 2011). Sonobuoy models classified as oceanic are shown as white squares, continental as black squares and transitional as blue diamonds. Dashed polygon indicates ocean-continent boundary (OCB; Table 2). The number of each solid blue line indicates the figure number of the seismic transects. D3 indicates a regional magnetic zone defined by Saltus et al. (2011). NoRMA: Northwind Ridge Magnetic Anomaly. Parallel white dots indicate a basement depression.

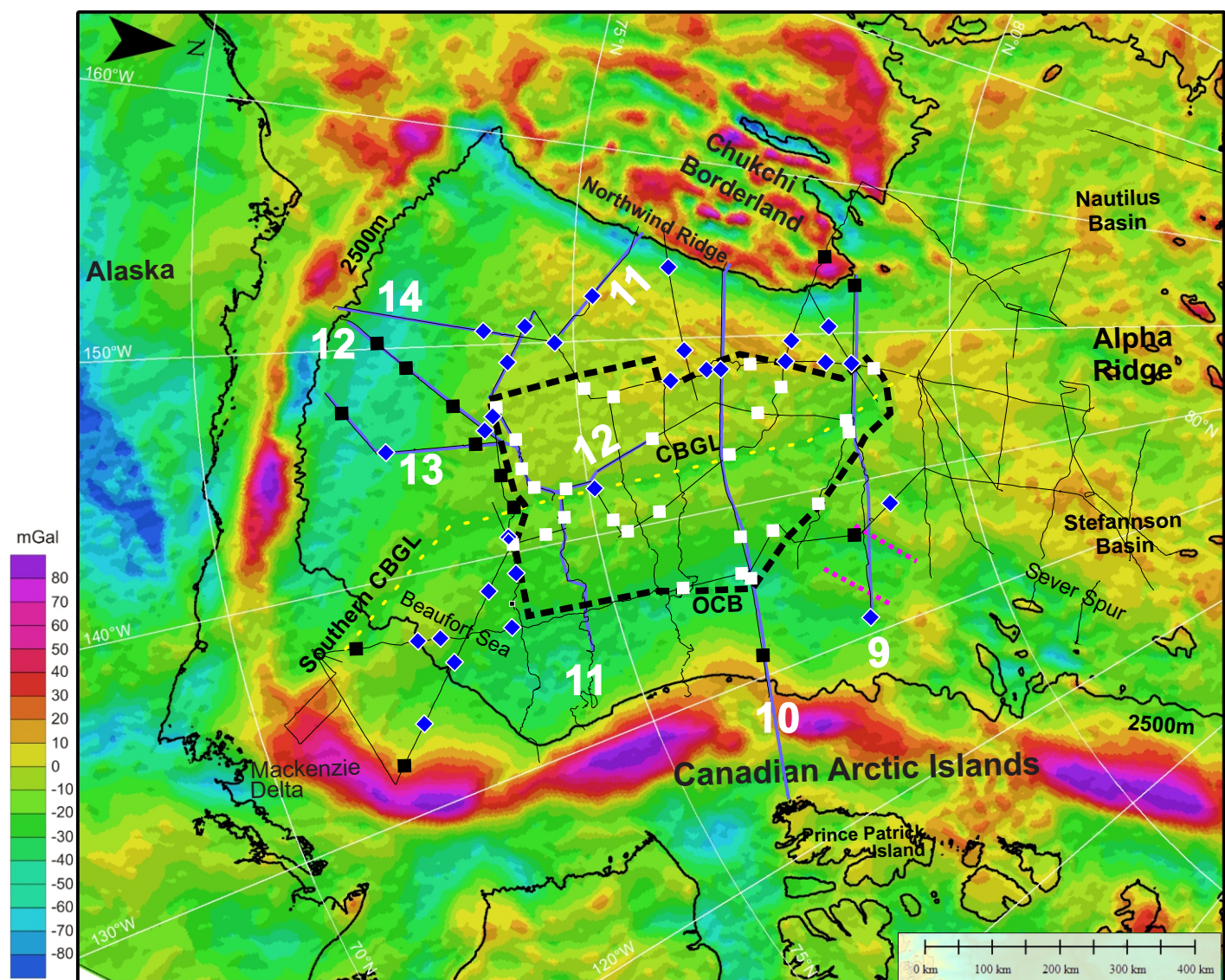
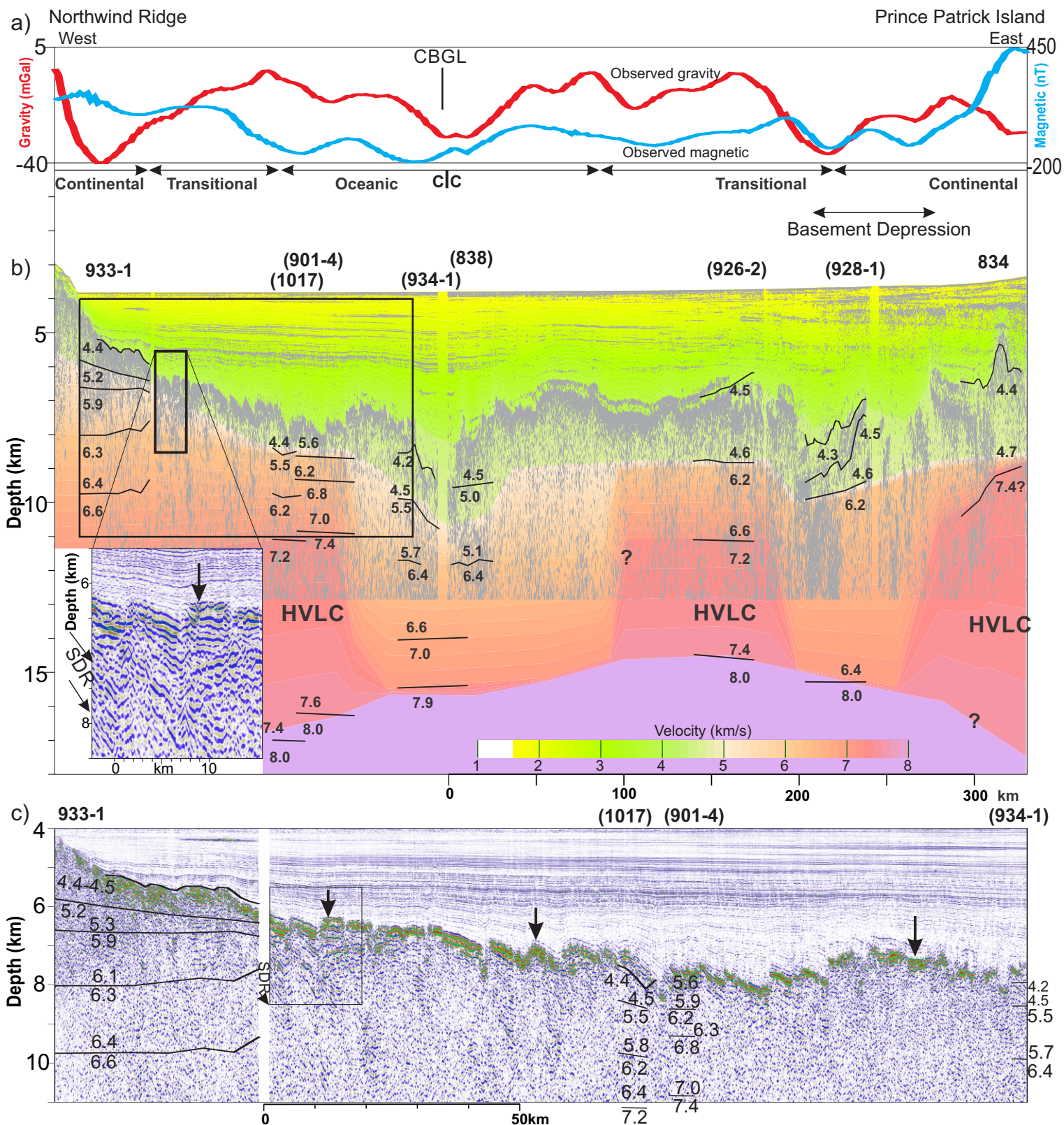


Fig. 8. Free Air gravity anomaly map of Canada Basin (with onshore Bouguer anomaly; after Andersen et al, 2010). Sonobuoy models classified as oceanic are shown as white squares, continental as black squares and transitional as blue diamonds. Dashed polygon indicates the ocean-continent boundary (OCB; Table 2). The number of each solid blue line indicates the figure number of the seismic transects. Parallel red dots indicate a basement depression.



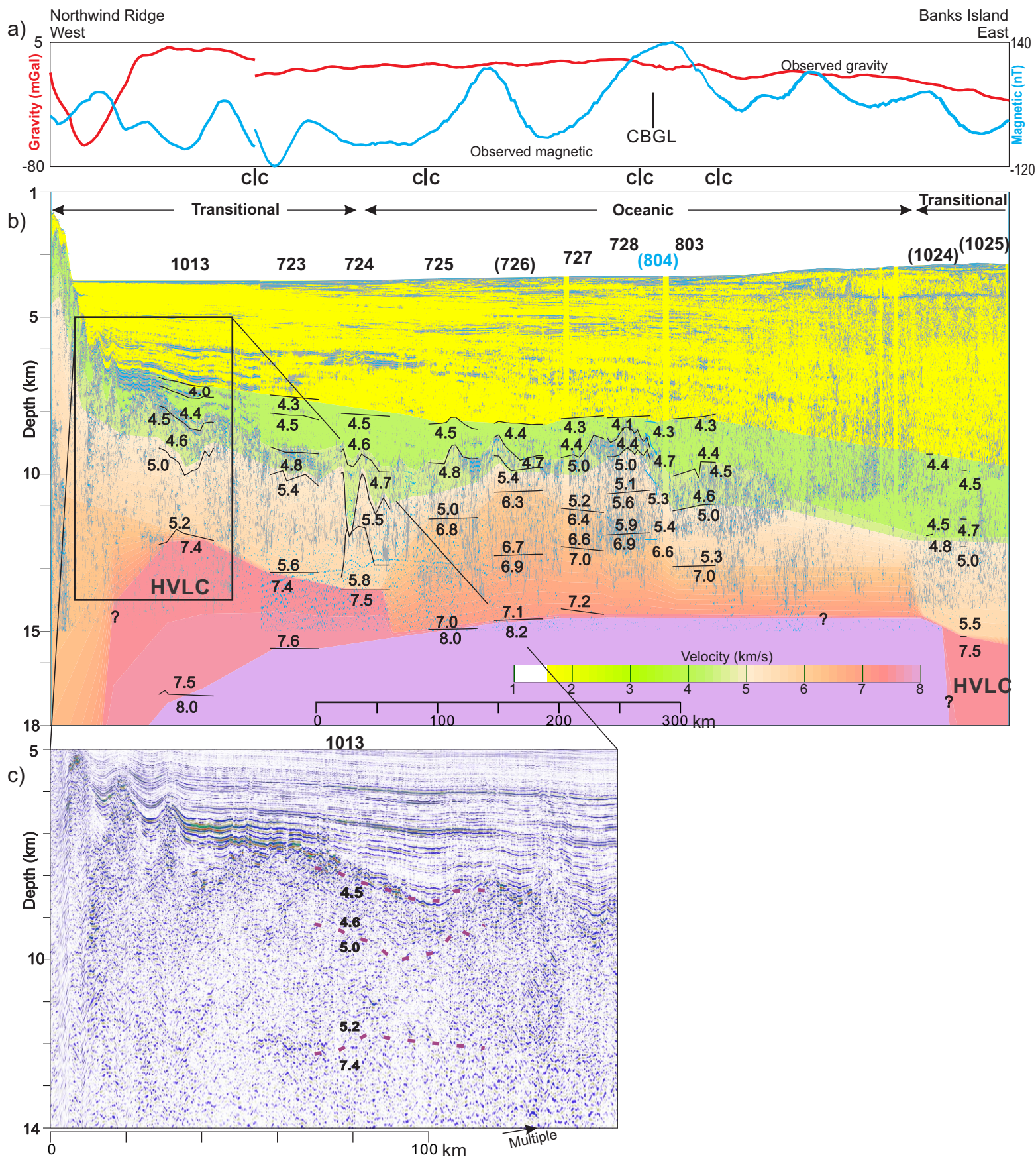


Fig. 11. Seismic transect across southern Canada Basin (see Fig. 1 for location). a) Magnetic (blue line) and gravity (red line) anomalies extracted from gridded data in Figs. 7 and 8. C|C indicates course changes. b) Velocity model overlain by depth-converted reflection profile. Sonobuoy 804 is indicated in blue to contrast with adjacent instruments. The rectangular area marked on panel b) is enlarged in c) to show details of deeper reflectors coincident with 7.4 km/s layer (vertical exaggeration: 11:1). See Fig. 9b for more description.

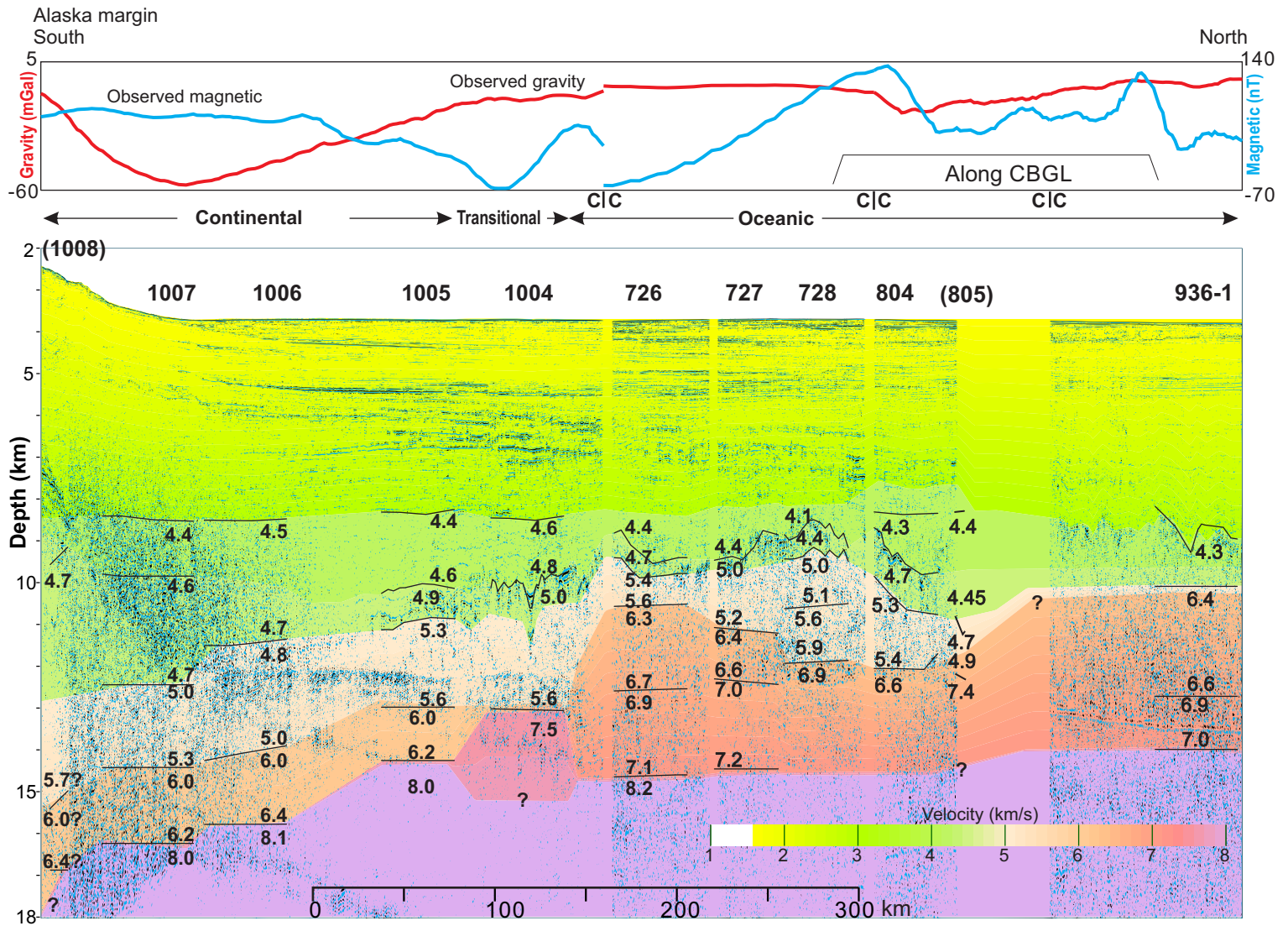


Fig. 12. Seismic transect from Alaskan margin northward to nearly parallel to Canada Basin Gravity Low (CBGL). Top: magnetic (blue line) and gravity (red line) anomalies extracted from gridded data in Figs. 7 and 8. Bottom: the transect combines 4 seismic profiles, crosses continental, transitional and oceanic crust, and shows distinctive features in crustal velocity structures as well as basement morphology. Reflection profiles are brute stack data converted to depth, with positive amplitudes plotted as grey and negative amplitudes as blue. See figs. 9a and 9b for more description.

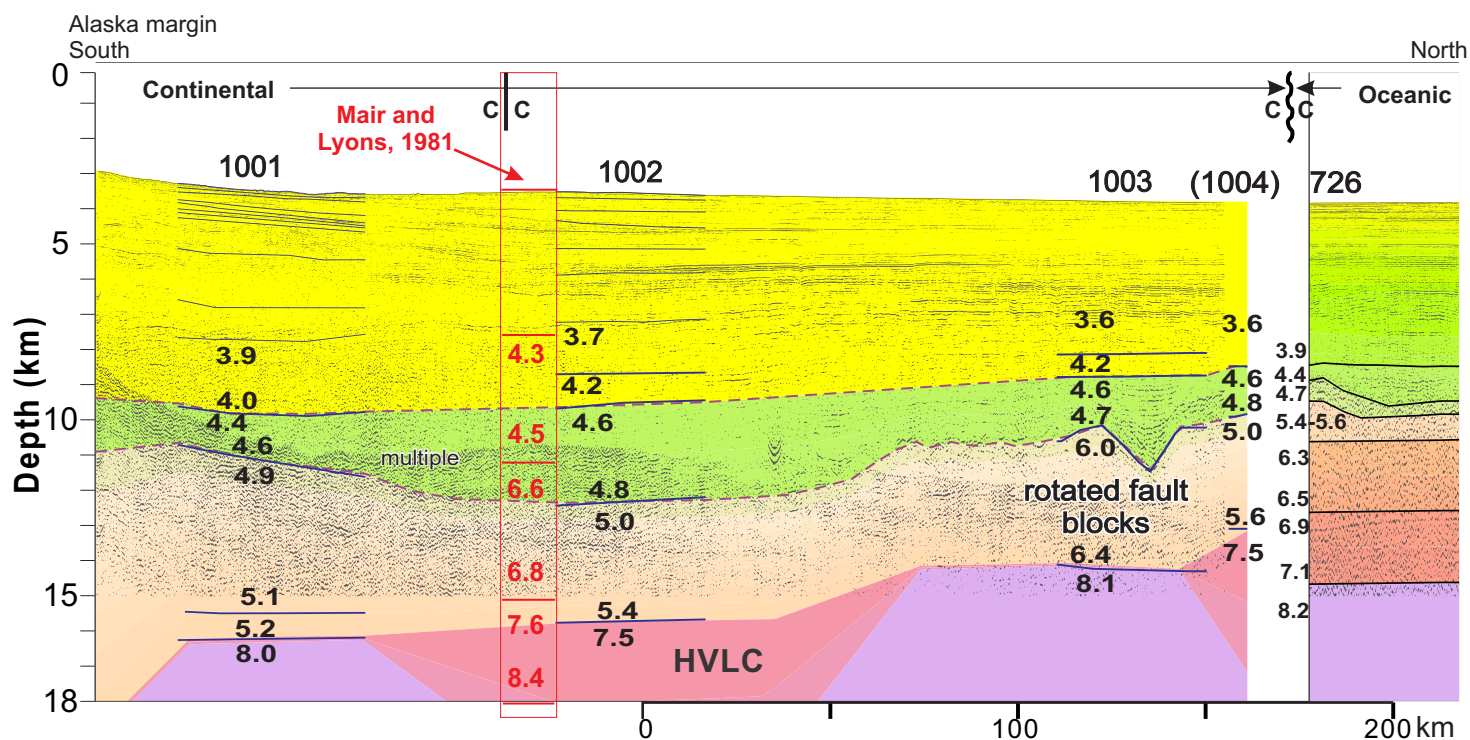


Fig. 13. Seismic reflection profile that overlaps the refraction lines reported in Mair & Lyons (1981). The velocities reported by them are plotted in red and the velocities based on our sonobuoy data are in black. Sonobuoy 1002 is categorized as in transitional zone due to an observation of HVLC. But this region is in general continental because: crustal velocity are in the continental range; crust shows consistent reflection characters with regional rotated fault blocks; and it is surrounded by typical continental crust in the vicinity of the shelf.

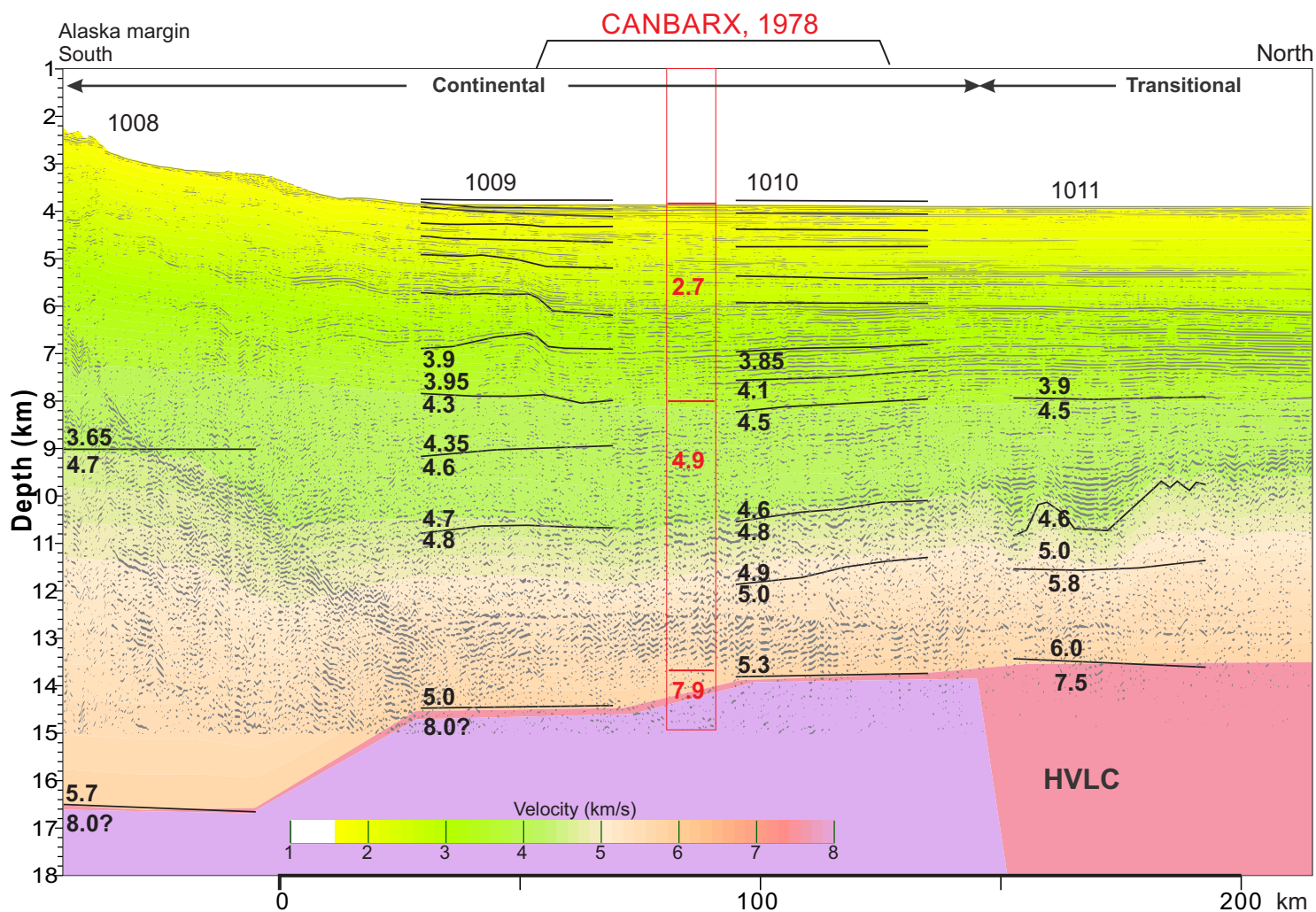


Fig. 14. Seismic reflection profile that overlaps the refraction lines reported in Baggeroer & Falconer (1982). The velocities reported by them are shown in red and labeled CANBARX 78.

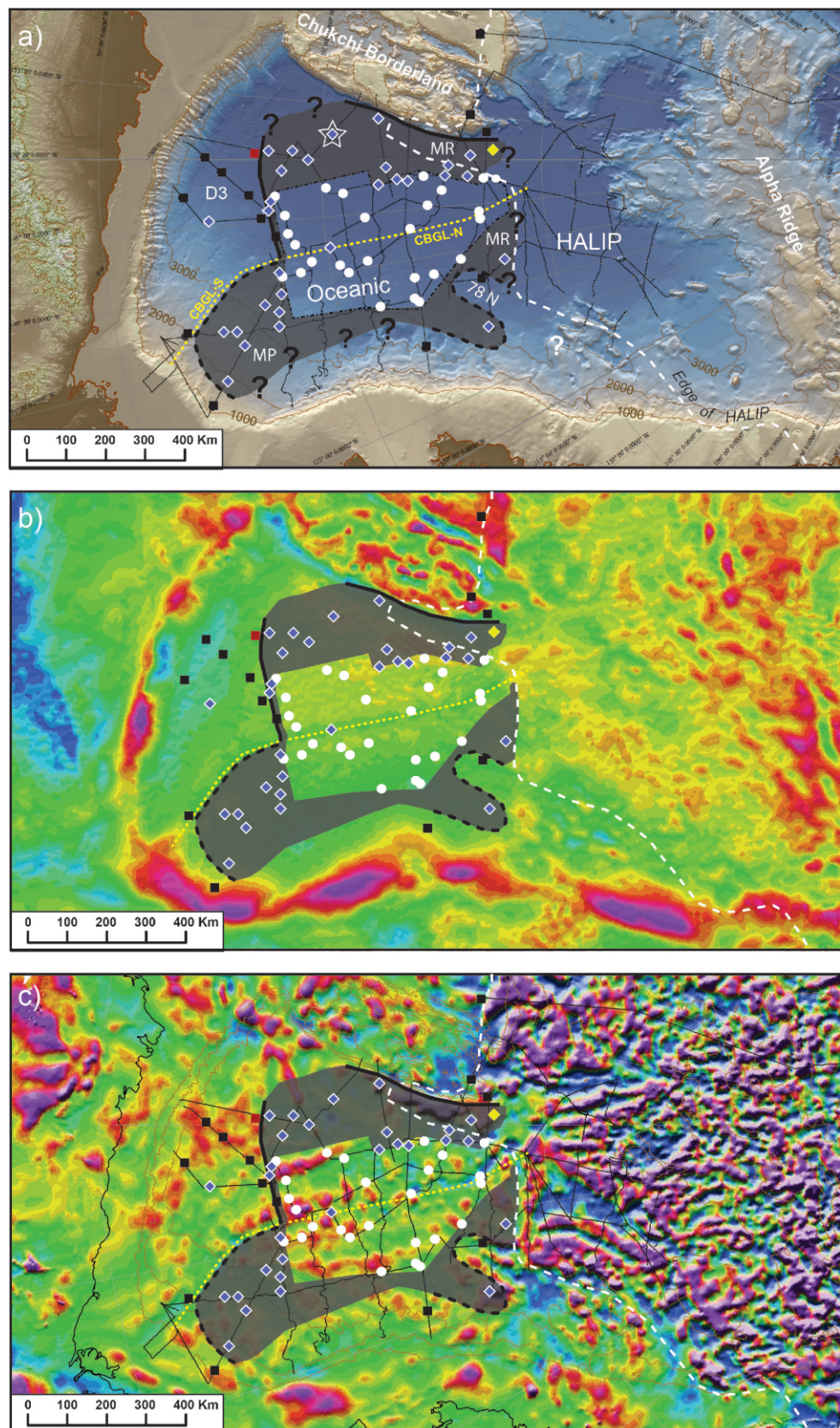


Fig. 15. Interpreted distribution of crustal regions plotted on a) bathymetric, b) gravity and c) magnetic maps (for colour bars of scale see Figs. 1, 8 and 7). The regions are based on the combination of sonobuoy velocities, magnetic and gravity data, seismic reflection control and geological context. The area with gray shading is the ocean-continent transition. Its inner limit is the ocean-continent boundary. The star shows the position of sonobuoy 1013. The red and yellow dots indicate the position of the Baggeroer & Falconer (1982) and Jackson et al. (1995) refraction experiments. Abbreviations: CBGL-N is Canada Basin Gravity Low-North, CBGL-S is Canada Basin Gravity Low-South, MR is Magma Rich, MP is Magma Poor, HALIP is High Arctic Large Igneous Province.